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ERRATA

Page 569, lines 14, 15, and 16, should read:

"geanticline do not move at the same rate, and the upper parts which were originally above the downward-moving secondary geosyncline may in a later stage of evolution be above the"

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Because of the high cost of printing, this Journal has found it necessary to adopt the following rules: ¶ 1. If articles exceeding 25 pages in length are accepted for publication in the Journal, the cost of composition *beyond 25 pages* shall be paid by the author or by the institution which he represents, at the current rates of the University Press. ¶ 2. The cost of illustrative matter, in excess of $1.00 per page for the article, shall be borne by the author. ¶ 3. While the cost of printing remains as now, the Journal cannot supply reprints without charge, but reprints will be furnished authors at cost, if ordered in advance of publication.

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Among articles to appear in early numbers of the *Journal of Geology* are the following:

Diastrophism and the Formative Processes. XIV. Megadiastrophism. By T. C. Chamberlin and R. T. Chamberlin.

The Physical Chemistry of the Crystallization and Magmatic Differentiation of Igneous Rocks. By J. H. L. Vogt.

The Mineralography of the Feldspars. By H. L. Alling.

Volcanic Earthquakes. By Charles Davison.

The Stratigraphic and Faunal Relationships of the Meganos Group, Middle Eocene of California. By Bruce L. Clark.

Types of Rocky Mountain Structure in Southeastern Idaho. By George Rogers Mansfield.

Summaries of Pre-Cambrian Literature of North America. Papers VI and VII. By Edward Steidtmann.

Russell Fork Fault of Southwest Virginia. By Chester K. Wentworth.

Cycles of Erosion in the Piedmont Province of Pennsylvania. By F. Bascom.

Through the generosity of an associate editor of the *Journal*, Dr. R. A. J. Penrose, Jr., the announced reduction to six issues will not be necessary for Volume XXIX. Subscribers will receive eight issues without change in price. (See Editorial Note in this number.)

VOLUME XXIX NUMBER 1

THE

JOURNAL OF GEOLOGY

JANUARY-FEBRUARY 1921

THE MECHANICAL INTERPRETATION OF JOINTS

WALTER H. BUCHER
University of Cincinnati

PART II

OUTLINE

I. MOHR'S THEORY OF RUPTURE[1]

BY LOUIS BRAND

Let P denote a point of a body at which the state of stress is to be investigated. For this purpose imagine a sphere of infinitesimal radius described about P as center (Fig. 10); then the totality of the stresses at all the surface elements of the sphere constitute the state of stress at the point P. A surface element of the sphere may be specified by means of the vector radius to the element. The surface r thus means an infinitely small plane surface tangent to the sphere at the end of r. In the following, the stresses shown

[1] This account of Mohr's theory of rupture was written at the writer's request by Professor Louis Brand, of the University of Cincinnati, to whom he wishes herewith to express his sincere thanks.

acting upon the sphere are those due to outside matter. Since we assume that the stress distribution within the body is continuous, the stresses that correspond to the two surface elements at the end of a diameter of the sphere are therefore numerically equal and opposite in sign, except for negligibly small quantities.

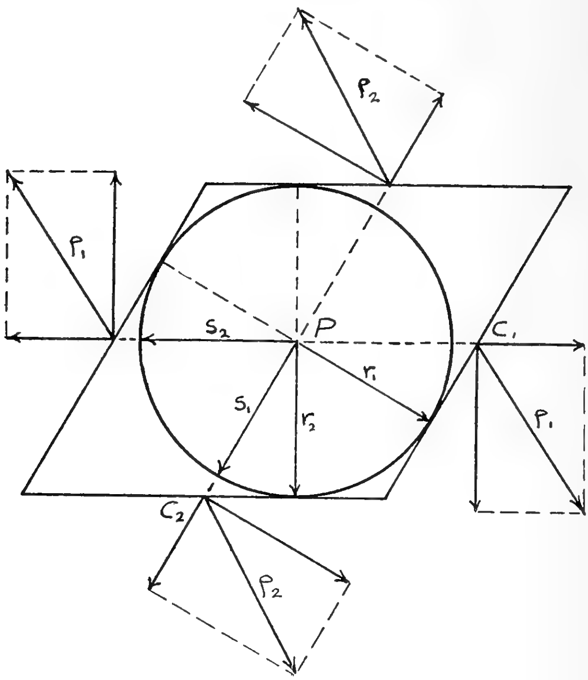

FIG. 10

We shall now obtain the relation between the unit stresses ρ_1, ρ_2 at the two surfaces r_1, r_2. Form a parallelepiped with three pairs of planes tangent to the sphere, one pair normal to r_1, another to r_2, while a third pair is normal to a radius r_3 which is perpendicular to both r_1 and r_2. Figure 10 shows the projection of this parallelepiped upon a plane through P normal to r_3. Both faces normal to r_3 project into a rhombus. The other four faces are rectangles of equal area, dA, and project into lines. Now resolve the stresses ρ_1 into three components: parallel to r_2, parallel to r_3, and normal

to r_2, r_3. Similarly resolve the stresses ρ_2 into three components: parallel to r_1, parallel to r_3, and normal to r_1, r_3. Of these components, only the first named of each set, namely,

$$\rho_1 \cos(\rho_1 r_2),\ \rho_2 \cos(\rho_2 r_1),$$

have moments about r_3. Hence for equilibrium as regards rotation about r_3, we have

$$2dA \cdot \rho_1 \cos(\rho_1 r_2) \cdot \overline{PC_1} = 2dA \cdot \rho_2 \cos(\rho_2 r_1) \cdot \overline{PC_2};$$

or, since $\overline{PC_1} = \overline{PC_2}$,

$$\rho_1 \cos(\rho_1 r_2) = \rho_2 \cos(\rho_2 r_1). \tag{1}$$

The angles indicated are those between the stresses and *outwardly directed radii*. The fundamental equation (1) may be stated as follows:

If ρ_1, ρ_2 are unit stresses at the surface elements r_1, r_2, the projection of ρ_1 upon r_2 is equal to the projection of ρ_2 upon r_1.

We proceed to put equation (1) in a more usable form. Let us regard angles in the plane of r_1, r_2 as positive when clockwise; and let s_1, s_2 denote radii in this plane 90° *ahead* of r_1, r_2 respectively. Now resolve ρ_1 into components as follows: (1) $\sigma_1 = \rho_1 \cos(\rho_1 r_1)$, parallel to r_1; (2) $\tau_1 = \rho_1 \cos(\rho_1 s_1)$, parallel to s_1; (3) parallel to r_3. Also resolve ρ_2 into the components: (1) $\sigma_2 = \rho_2 \cos(\rho_2 r_2)$, parallel to r_2; (2) $\tau_2 = \rho_2 \cos(\rho_2 s_2)$, parallel to s_2; (3) parallel to r_3. The *normal stresses* σ_1, σ_2 (i.e., normal to their surface elements) are positive when directed outward from P, or when they are *tensions*. The components τ_1, τ_2 of the *shearing stresses* are positive when they produce clockwise moments about r_3. We now transform the cosines in (1) by means of the formula for the angle between two directions in space:

$$\begin{aligned}
\cos(\rho_1 r_2) &= \cos(\rho_1 r_1)\cos(r_2 r_1) + \cos(\rho_1 s_1)\cos(r_2 s_1) + \cos(\rho_1 r_3)\cdot\cos(r_2 r_3) \\
&= \cos(\rho_1 r_1)\cos(r_1 r_2) + \cos(\rho_1 s_1)\sin(r_1 r_2), \\
\cos(\rho_2 r_1) &= \cos(\rho_2 r_2)\cos(r_1 r_2) + \cos(\rho_2 s_2)\cos(r_1 s_2) + \cos(\rho_2 r_3)\cos(r_1 r_3) \\
&= \cos(\rho_2 r_2)\cos(r_1 r_2) - \cos(\rho_2 s_2)\sin(r_1 r_2).
\end{aligned}$$

Substituting the values in (1) and noting the above values for σ_1, σ_2, τ_1, τ_2, we have

$$\sigma_1 \cos (r_1 r_2) + \tau_1 \sin (r_1 r_2) = \sigma_2 \cos (r_1 r_2) - \tau_2 \sin (r_1 r_2),$$

or

$$(\sigma_2 - \sigma_1) \cos (r_1 r_2) = (\tau_1 + \tau_2) \sin (r_1 r_2). \tag{2}$$

As a first application of this equation we have

$$\tau_1 + \tau_2 = 0 \quad \text{when} \quad (r_1 r_2) = 90^\circ. \tag{3}$$

In this case the shearing stresses are numerically equal but opposite in sign.

Next, let r_0 be any fixed radius in the plane r_1, r_2; and write

$$\phi = (r_0 r_1),\ \phi + \Delta\phi = (r_0 r_2);\ \ \sigma_2 = \sigma_1 + \Delta\sigma,\ \tau_2 = \tau_1 + \Delta\tau.$$

Then (2) assumes the form

$$\Delta\sigma \cos \Delta\phi = (2\tau_1 + \Delta\tau) \sin \Delta\phi;$$

dividing through by $\Delta\phi$ and letting $\Delta\phi$ approach zero, we obtain

$$\frac{d\sigma}{d\phi} = 2\tau, \tag{4}$$

the subscript being no longer needed. From this equation it appears that σ increases with ϕ when τ has the direction of increasing ϕ.

If σ is not constant over the sphere it must reach a minimum σ_x and a maximum σ_z at certain diameters which we denote by x and z respectively. At these points $\frac{d\sigma}{d\phi} = 0$ for *all* diametral planes; hence from (4) the *total* shear vanishes at x and z. Also, from (2), $\cos (xz) = 0$, so that x and z are perpendicular. Again, if y is a third diameter, perpendicular to both x and z, the components of shear in both yx and yz planes vanish by virtue of (3), and hence the *total* shear for y is also zero. The three mutually perpendicular diameters x, y, z, are called the *principal axes*, the planes xy, yz, zx, the *principal planes*, and the stresses σ_x, σ_y, σ_z, the *principal stresses*

at the point P. When the principal stresses are unequal the notation has been chosen so that $\sigma_x < \sigma_y < \sigma_z$.

We now choose positive directions on the principal axes, and denote the positive principal radii by r_x, r_y, r_z. Then, if ρ is the stress at any surface r, we have from (1)

$$\left.\begin{aligned} \rho \cos(\rho r_x) &= \sigma_x \cos(r r_x), \\ \rho \cos(\rho r_y) &= \sigma_y \cos(r r_y), \\ \rho \cos(\rho r_z) &= \sigma_z \cos(r r_z). \end{aligned}\right\} \qquad (5)$$

The components of the stress at any surface r are thus given in terms of the principal stresses and the direction cosines of r. From these components we may compute the normal stress, σ, at r:

$$\sigma = \rho \cos(\rho r) = \sigma_x \cos^2(r r_x) + \sigma_y \cos^2(r r_y) + \sigma_z \cos^2(r r_z). \qquad (6)$$

The equations (5) show, moreover, that the stresses over the sphere are symmetrically distributed with respect to each of the three principal planes. It is sufficient, therefore, to know the stress distribution in one of the octants into which the principal planes divide the sphere.

We shall now examine the stress distribution in the principal planes. Let r be a radius in the xz plane making an angle $\phi = (r r_z)$ with the positive z-axis. Since $(r r_y) = 90°$, we have from (5) that $(\rho r_y) = 90°$; hence the stresses in a principal plane lie entirely in that plane. From (6) we have for the normal stress at r

$$\sigma = \sigma_x \sin^2\phi + \sigma_z \cos^2\phi = \frac{\sigma_x + \sigma_z}{2} + \frac{\sigma_z - \sigma_x}{2} \cos 2\phi. \qquad (7)$$

To obtain τ we put $\sigma_2 = \sigma_z$, $\tau_2 = \tau_z = 0$, $r_2 = r_z$ in (2); then, by virtue of (7),

$$\tau = (\sigma_z - \sigma) \frac{\cos\phi}{\sin\phi} = (\sigma_z - \sigma_x) \sin\phi \cos\phi = \frac{\sigma_z - \sigma_x}{2} \sin 2\phi. \qquad (8)$$

If σ, τ are regarded as rectangular co-ordinates, equations (7) and (8) are the parametric equations of a circle of center $[(\sigma_x + \sigma_z)/2, 0]$ and radius $(\sigma_z - \sigma_x)/2$. We thus have the following graphical representation of the stress distribution in the xz plane: Lay off

$\overline{OX}=\sigma_x$, $\overline{OZ}=\sigma_z$ on the axis of abscissas (Fig. 11), and draw a circle having $\overline{XZ}$ as diameter. This circle has the center and radius specified above; and the abscissa and ordinate of any point R of the circle represent the normal and shearing stress, σ, τ, respectively, which correspond to a radius r of the sphere in the xz plane and making an angle ϕ with r_z equal to one-half of the central angle ZCR.

If we lay off $\overline{OY}=\sigma_y$, a similar argument shows that the stress distributions on the xy and yz planes are represented by the co-ordinates of the points forming the circles on $\overline{XY}$ and $\overline{YZ}$ as

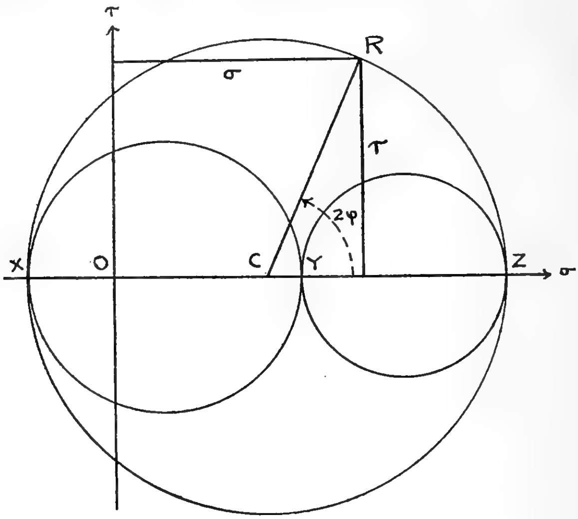

FIG. 11

diameters. It may also be shown that for any radius of the sphere not in the principal planes, the corresponding stresses σ, τ, are represented by the co-ordinates of points within the region bounded by all three of the circles XY, YZ, XZ. *The circle XZ therefore passes through all the points which have the greatest τ for any given σ occurring in the stress distribution.* This circle plays an important rôle in Mohr's theory of rupture, and is called by Mohr the *principal circle* in the graphic representation of the state of stress.

Mohr's theory of rupture deals with cases in which the failure of the material is assumed to be due to the sliding of one layer over another in certain planes called *shearing planes.* The fundamental

postulate of this theory may be stated as follows: *The ultimate strength of the material is determined by the greatest shearing stress in the shearing planes, this limiting shear itself depending upon the normal stress in the planes.* The limiting shear τ_m is thus assumed to be some function of the normal stress in the plane, $\tau_m = f(\sigma)$, but the nature of this dependence is not specified by the theory. The curve, however, which represents the relation $\tau_m = f(\sigma)$ has certain properties which may be inferred from Mohr's graphic representation of stress distributions. In this representation a point

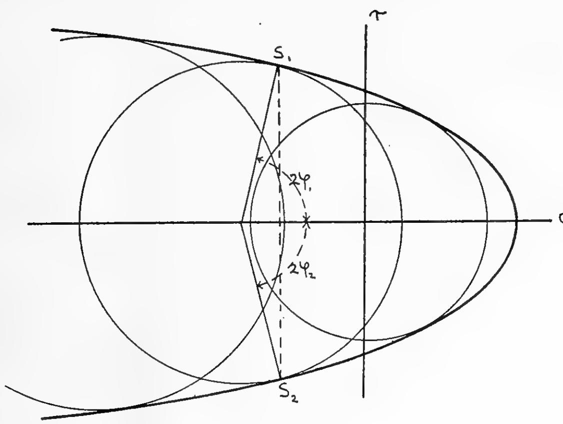

FIG. 12

S_1, corresponding to a shearing plane (Fig. 12), must lie on a principal circle, since the points of this circle have the greatest τ for a given value of σ. From this fact we may draw two important conclusions:

1. Since a principal circle is completely determined by the two extreme principal stresses, σ_x, σ_z, the ultimate strength of a material that fails by slippage must be entirely independent of the intermediate principal stress σ_y.

2. Since points on a principal circle correspond to planes normal to the xz plane, the shearing planes at any point must pass through the y-axis.

Furthermore, a point on a principal circle corresponding to a shearing plane must lie on the envelope of all the principal circles

representing ultimate states of stress. For this envelope is the locus of all points, which, for a given σ, have the greatest τ for the stress distributions in question. The envelope thus represents the relation $\tau_m = f(\sigma)$ between maximum shear and normal stress. As all principal circles have their centers on the σ-axis, their envelope is symmetric with respect to this axis, and touches each circle in two points, S_1, S_2, corresponding to the two shearing planes for the state of stress represented by the circle. Referring to Figure 12, we see that the shearing planes corresponding to S_1 and S_2 make equal angles with the zy plane: $\phi_1 = \phi_2$. The acute angle between the planes is therefore equal to $2\phi_1$, and is bisected by the principal plane zy.

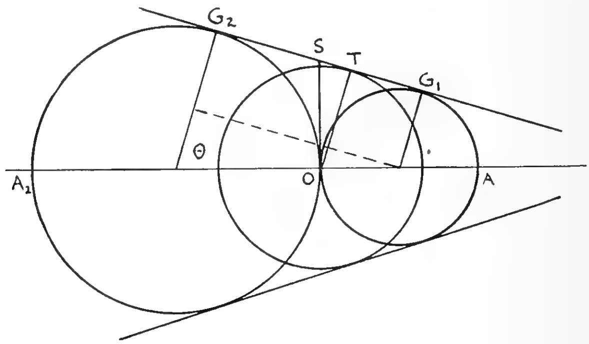

FIG. 13

Experiment has shown that the envelopes are roughly parabolic in shape. They presumably cut the $+\sigma$-axis at the point-circle representing rupture due to uniform all-sided tension ($\sigma_x = \sigma_y = \sigma_z$). Mohr also infers from the apparent impossibility of crushing a substance by applying uniform all-sided pressure, that if the envelope cuts the $-\sigma$-axis at all, it does so at an excessive distance from the origin.

If the ultimate tensile and compressive strength of a material are known, say σ_1 and σ_2, we may construct the principal circles corresponding to these limiting states, namely,

$$\sigma_x = \sigma_y = 0, \quad \sigma_z = \sigma_1; \qquad \sigma_x = -\sigma_2, \quad \sigma_y = \sigma_z = 0.$$

Thus in Fig. 13 we lay off $\overline{OA}_1=\sigma_1$, $\overline{OA}_2=-\sigma_2$, and draw the principal circles on $\overline{OA}_1$ and $\overline{OA}_2$. For states in which the principal stresses do not considerably exceed σ_1 and $-\sigma_2$, we may regard the common external tangents to these circles as representing approximately a portion of the envelope. For the region in which this approximation is allowable, the angle between the shearing planes is constant and given by

$$\cos\theta=\frac{\sigma_2-\sigma_1}{\sigma_2+\sigma_1}=\frac{\kappa-1}{\kappa+1}, \quad \kappa=\frac{\sigma_2}{\sigma_1}. \tag{9}$$

The angle θ is thus independent of the particular state of stress and depends only upon the ratio of ultimate compressive to ultimate tensile stress. According as $\sigma_2 > \; = \; < \sigma_1$, θ will be an acute, right, or obtuse angle. For most materials $\sigma_2>\sigma_1$; values of θ corresponding to different values of κ in this case are given in the following table:

$\kappa=\frac{\sigma_2}{\sigma_1}$	1	1.5	2	3	4	5	10	20
θ	90°	78°	71°	60°	53°	49°	35°	25°

The principal circle for the state in which $\sigma_x=-\sigma_z$, that is, a circle about O as center and tangent to the envelope tangents, determines the ultimate torsional strength: $\rho_3=\overline{OT}$. The ultimate shearing strength is given by $\tau_4=\overline{OS}$; for this ordinate represents the maximum shear for zero normal stress. From the geometry of the figure,

$$\rho_3=\frac{\sigma_1\sigma_2}{\sigma_1+\sigma_2}, \quad \tau_4=\tfrac{1}{2}\sqrt{\sigma_1\sigma_2}\,.$$

When σ_1 and σ_2 are nearly equal we have $\rho_3=\tau_4=\frac{1}{2}\sigma$, a result in concordance with many old and modern experiments and not satisfactorily explained by the earlier theories of rupture.

II. MOHR'S THEORY APPLIED TO EXPERIMENTAL DATA

Mohr's formula, $\cos\theta=\frac{\sigma_2-\sigma_1}{\sigma_2+\sigma_1}$, connects the angle of shearing with two important physical constants, the ultimate tensile strength and the crushing strength.

Within certain limits, that is to say, in so far as the curve $\tau_m=f(\sigma)$ between the points G_1 and G_2 approaches a straight line, it confirms

Hartmann's observation that the angle of shearing is independent of the nature and intensity of the stresses involved.

It also brings out clearly, in a qualitative way, the very different attitude of the planes of shearing in brittle and ductile substances. This becomes apparent when we introduce into it the average values of ultimate strengths contained in the following table:

TABLE I*

	k_2 Tons	k_1 Square Inches	k_2/k_1
Granite (av.)	7.5	0.3	25
Glass (various sorts)	4.3–8.7	2.5–6.0	15+
Marble (av.)	4.0	0.35	11
†Bluestone	6.8	0.7	9.7
Limestone (av.)	3.5	0.5	7
Glass (common green)	10	1.5	6.7
Iron, cast	40	7.5	5.3
Zinc, cast	(10)	2.5	(4+)
Georgia Yellow Pine (across grain)	0.7	0.3	2.3
Copper, cast	(20)	12	1.7
Steel, cast	35	35	1
Copper, bolts	15	15	1
Slate	5	5	1
White Oak (across grain)	1	1	1
White Oak (with grain)	3.5	5	0.7
Iron, wrought (av. good bars)	16–20	25	0.6–0.8
Georgia Yellow Pine (with grain)	2.5	6	0.4

*All figures in this table, excepting those in the third row, from H. H. Suplee, *The Mechanical Engineer's Reference Book* (4th ed., 1913).

They represent at best only average values and are used here only in a qualitative sense. Values of K_2 given in parenthesis indicate loads producing 10 per cent reduction in original lengths.

All figures are given in the original in lbs. per sq. in.

The values for various sorts of glass are taken from Winkelmann and Schott, quoted in O. D. Chwolson, *Lehrbuch der Physik*, Vol. I (Braunschweig, 1902), pp. 709 and 712.

† "Fine-grained, compact, very strong and durable" graywacke (of Hamilton age). G. P. Merrill, *Stones for Building and Decoration* (2d ed., 1897), p. 322.

According to the formula, the angle θ will differ the more from 90° the greater the ratio of the values of crushing and tensile strength.

Correspondingly, we find that substances like glass shear at very acute angles,[1] while for mild steel the angle varies between 80° and 100°.[2]

[1] 20°–30° in the case of glass of thick microscopic slides, according to tests made by the writer.

[2] See results of experiments by W. Mason and G. H. Gulliver, as given in W. Mason, "The Lüders' Lines on Mild Steel," *Proc. Phys. Soc. of London*, Vol. XXIII (1911), p. 306.

In the case of those substances in which the tensile strength is greater than the crushing strength, the formula would indicate that the shearing angle is obtuse. That this is indeed the case, for instance in the striking case of wood cut with the grain, may be seen from Figure 3, page 17, of Leith's *Structural Geology*.

This leads us to the important generalization that the angle of shearing of a material is the more acute the more brittle the substance is, and vice versa. In fact, it seems possible, if not probable, that in the hands of the physicist the angle of shearing will be made the chief criterion of brittleness.

The formula also brings out clearly the fact that the angle of shearing is independent of the hardness of the material. In the table given above wrought iron ranks with oak and pine wood. Small cubes of a "brittle" rubber, sold under the trade-name "soap rubber," produce shearing planes in the form of pyramids exactly like those seen in cubes of sandstone in ordinary crushing tests, with apical angles of 50° or even less. This shows that the shearing angle is also independent of the absolute amount of deformation of which the substance is capable below the elastic limit.

Strongly ductile[1] bodies, on the other hand, like soap, shear at very obtuse angles.

[1] The writer knows that in this paper he is using the word "ductile" in a sense which is sure to be severely criticized. He would be delighted to see such criticism lead to a fruitful open discussion of the fundamental conceptions involved in the deformation of solids. At this place the restricted sense in which the word "ductility" is used here may be defined best by giving it as one of several purely empirical characteristics of solids under deformation.

Substances differ in

1. The force required to produce the same absolute amount of deformation. (Small: rubber, wet clay; large: steel.)

2. The absolute amount of deformation required to reach the elastic limit. (Small: steel, wet clay; large: rubber.)

3. The percentage of any given deformation which remains permanent when the stress is removed. (0 per cent = perfect elasticity; 100 per cent = perfect plasticity.)

4. The additional force required to produce an additional amount of permanent deformation (negative, zero, positive).

5. The time required to produce the same absolute amount of permanent deformation without rupture.

6. The position of the point of rupture with reference to the yield point. (Point of rupture <or> yield point.)

In this paper a substance is called "brittle" when its point of rupture lies near its yield point. It is called "the more ductile" the farther beyond the yield point its point of rupture lies. In a "perfectly ductile" substance it lies an infinite distance beyond. In a "perfectly brittle" substance it is reached before the yield point.

When rocks are subjected to deformation in nature, the question whether they will break or bend depends largely on the degree of brittleness they possess under the given conditions and not on their hardness. This explains why geologists have been justified in the attempt to reproduce in the laboratory the various structures exhibited by the hard materials of the earth's crust by the use of soft clay, mixtures of clay and plaster of paris, and even wet sand. The use of the angle of shearing as an index of brittleness opens up new possibilities for standardizing the materials used in such experiments to accurately reproduce definite actual conditions.[1]

Mohr's formula could be quantitatively correct only, if the curve $\tau_m=f(\sigma)$ were practically a straight line between the circles of ultimate tensile and compressive stress. This, however, is not true. The form of the curve, therefore, must first be determined experimentally for each substance. From it the variable θ can be computed, from point to point by analogous formulas.

This was carried out for the first time, as far as the writer knows, in a series of excellent experiments by Kármán.[2]

He used an apparatus in which small cylinders of rock could be subjected simultaneously to hoop and longitudinal pressures in such a way that either pressure could be controlled without changing the other. The results of his experiments are embodied in stress-strain diagrams which in the most striking way show the fact that the materials used—marble and sandstone—change step by step with increasing hoop pressure from a state of perfect brittleness to one of perfect ductility. In this respect Kármán's experiments supplement beautifully the brilliant investigations of Adams.

With low hoop pressures shearing occurred in the rock cylinders resulting in the formation of Lüders' lines on the polished surfaces and, with lowest hoop pressures, leading to rupture.

In the following table[3] the observed values of the angles of shearing at various hoop pressures are placed side by side with the values computed according to Mohr's graphic construction.

[1] The angle of shearing can readily be determined for many substances by means of an ordinary vise, if small cubes (1 cm^3) are used.

[2] Th. von Kármán, "Festigkeitsversuche unter allseitigem Druck," *Zeitschr. des Vereins deutscher Ingenieure*, Vol. LV (1911), pp. 1749–57.

[3] Tables 1 to 4 of Kármán's paper.

The agreement is sufficiently close to strongly support Mohr's theory. In addition, however, the figures reveal the striking fact that as the material changes, under the action of all-sided pressure, from a brittle to a ductile substance, the angle of shearing grows progressively less and less acute. Kármân's experiments have, therefore, completely verified in the case of one and the same substance the inference that the less brittle a substance is the larger is its angle of shearing.

TABLE II

	Hoop Pressure $\sigma_2=\sigma_3$ in Atmospheres	Effective Longitudinal Pressure $\sigma_1-\sigma_2$ in Atmospheres	θ Observed without Reduction*	θ Observed with Reduction	θ Computed
Marble	0	1360	54°	54°	53°
	235	2100	59°	58°	58°
	500	2650	72°	65°	63°
	685	2880	83°	70°	73°
Sandstone	0	690	38°	38°	40°
	280	2040	70°	69°	63°
	555	2580	82°	73°	70°

* To find the true angle at which rupture actually took place, it is necessary to reduce the observed angle to the value it had when the rock cylinder was deformed under its load.

We can, however, go even one step farther. If growing circumferential pressure at right angles to the direction of maximum (compressive) stress increases the ductility of a substance, and with it the angle of shearing, circumferential tension must decrease it, that is, render the substance more brittle. This is completely born out by the experiments published in 1911 by W. Mason.[1] He subjected tubes of mild steel to longitudinal compressive stress simultaneously with the application of interior hydrostatic pressure, and in one series of experiments, made the tubes undergo longitudinal tension while applying water pressure externally. The angle facing the direction of maximum compressive stress, in the absence of circumferential pressure, measured[2] approximately 100°. With growing tension normal to the direction of compression,

[1] W. Mason, "The Lüders' Lines on Mild Steel," *Proc. Phys. Soc. of London*, Vol. XXIII (1911), pp. 305–33.

[2] *Ibid.*, Table B, values at bottom of column.

the angle changed from 100° to values as low as 84° and even[1] 79°. These values remained the same, whether the compressive stress acted longitudinally or as hoop stress.

Here, then, we have a substance which normally shears under compression at an angle of 100° changed to one shearing at 80° through the action of tension in all directions normal to that of the axial compression, or, using the word in the sense defined above, we may say, the substance has been made more brittle.

III. THE ELLIPSOID OF STRAIN

The attempt to correlate joint planes with stress-strain relations, to which the first part of this paper was devoted exclusively, is by no means new. Steidtman's splendid paper on "The Secondary Structures of the Eastern Part of the Baraboo Quartzite Range, Wisconsin"[2] is well known, and Leith, in his lectures and in his *Structural Geology*[3] has impressed on the younger generation of geologists the importance of shearing planes in the mechanics of rock fractures by the use of a wire-netting model.

Unfortunately, however, the model as well as the earlier discussions by other writers, give expression only to that case in which the planes of shearing form an obtuse angle in the direction of compressive stress.

The outstanding characteristic of the strain ellipsoid illustrated by Leith's wire-netting model, is the fact that the elongation in the direction of one principal stress equals the shortening in the direction of the other principal stress, or, that the area of the strained surface remains unchanged. A simple mathematical consideration shows that when a circle is changed into an ellipse of equal area, the angle of the lines of no distortion facing the direction of shortening, must always exceed 90°[3] (Fig. 14). To reduce this angle to the smaller value characteristic of all brittle materials, we must assume the longitudinal shortening to be smaller than

[1] W. Mason, "The Lüders' Lines on Mild Steel," *Proc. Phys. Soc. of London*, Vol. XXIII (1911), Table D, bottom of column (complementary angle).

[2] *Jour. Geol.*, Vol. XVIII (1910), pp. 259–70.

[3] New York: Henry Holt & Co., 1913, pp. 18–20.

the transverse elongation (Fig. 15), that is, increase the area of the Figure under deformation.[1]

This, however, leads to the conclusion that under simple non-rotational stress a brittle body suffers an increase of volume. Since the angle of shearing is the more acute the more brittle a substance is, we must expect the increase of volume under stress to be the greater the more brittle a material is.

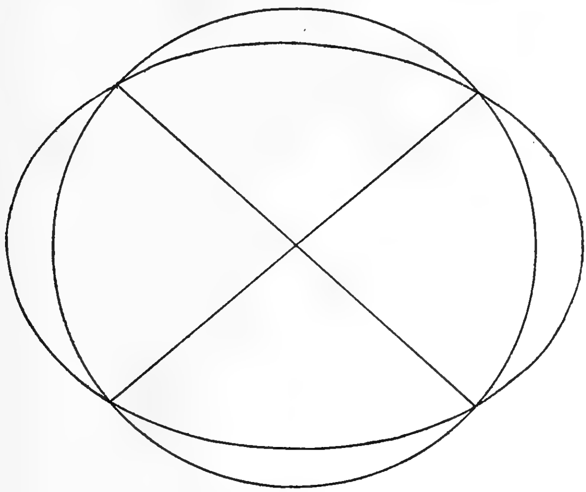

FIG. 14.—Diagram showing the position of the lines of no distortion in an ellipse of strain derived from a circle of equal volume. The angle of shearing is obtuse.

This seems indeed to be the case. Chwolson[2] gives the following formula connecting the modulus of volume increase (under tension), η, with the modulus of longitudinal strain (Young's modulus), a, and Poisson's ratio $\left(\frac{\text{lateral strain}}{\text{longitudinal strain}}\right)$, σ:

$$\eta = a(1-2\sigma).$$

According to this formula, $\sigma = 0.5$, when $\eta = 0$, that is, when the volume remains unchanged during deformation. The change of

[1] For the mathematical proof of this statement the writer is again under obligation to Dr. Brand.

[2] O. D. Chwolson, *Lehrbuch der Physik*, Vol. I, p. 713.

volume, therefore, must be the greater the more σ differs from 0.5, that is, the smaller it is. The following values[1] of σ for different substances completely confirm the inference drawn from the graphic representation of the strain ellipsoid.

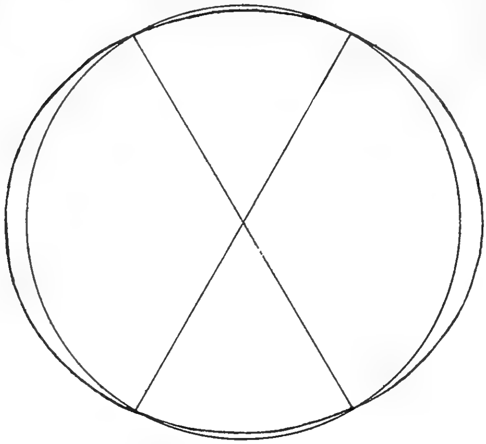

FIG. 15.—Diagram showing the position of the lines of no distortion in an ellipse derived from a circle of smaller volume (increase of volume accompanies deformation). The angle of shearing is acute.

Paraffin	0.50
Caoutchouc	0.50
Copper	0.348
Mild Steel	0.304
Iron	0.243 to 0.310
Zinc	0.205
Glass	0.197 to 0.319

That brittle bodies suffer an increase of volume when deformed under tension, is well known. That such an increase of volume also actually accompanies deformation under one-sided compression, as demanded by the graphic construction of the strain ellipsoid, seems to be proved by the experiments made by Kahlbaum and Seidler[2] and more recently by Lea and Thomas.[3]

It is essential, therefore, before we use the strain ellipsoid for the interpretation of shearing planes in nature, that we decide which form of the ellipsoid corresponds to the conditions of the specific case.

IV. PLANES OF SHEARING PRODUCED BY IRROTATIONAL AND ROTATIONAL STRAINS

We may now return to the interpretation of planes of shearing observed in nature. We have learned that Hartmann's law applies to brittle substances only, that is, that only in brittle materials the

[1] O. D. Chwolson, *Lehrbuch der Physik*, Vol. I, p. 714.

[2] R. Kahlbaum and Seidler, *Zeitschr. Anorg. Chem.* (1902), pp. 29–30, 254–94.

[3] F. C. Lea and W. N. Thomas, "Change in Density of Mild Steel Strained by Compression beyond the Yield-Point," *Engineering*, Vol. C (1915), pp. 1–3.

acute angle of shearing planes faces the direction of the compressive stress. We may now extend the law by adding, that in ductile substances it is the obtuse angle that faces the direction of the compressive stress.

Before attempting to apply the law to any specific case, therefore, we must decide whether the material under the given conditions had the properties of a brittle or those of a ductile substance. On the other hand, when the direction of the greatest principal stress is known, the position of the joint planes produced by it may be used to determine the degree of ductility which the material possessed at the time of shearing.

All the cases so far discussed involve irrotational strains only. The arrangement of shearing planes due to rotational strains, as illustrated by Leith's wire-netting model and discussed in his book on *Structural Geology*, is, of course, only possible in ductile substances, as a glance at the angle of the shearing planes will show. This model has, however, been applied successfully to some striking cases of jointing in quartzites.

We may approach the problem involved in these interesting cases by turning to an illustration in Van Hise's "Principles of North American Pre-Cambrian Geology," page 652.[1] Figure 131 shows layers of quartzite alternating with thin beds of more slaty character. The harder beds are traversed by two systems of intersecting joints, both forming angles of 50°–70° with the bedding planes, that is, forming acute angles of approximately 60° facing the bedding planes.

In the intercalated slaty beds, however, only one of these two joint systems is developed. It consists of more numerous joints inclined but 20° or less to the bedding planes. If the complementary symmetrical set were developed, the angle formed by the two systems facing the bedding planes would be 130° in these less brittle slaty beds, instead of 60° as in the brittle purer beds.

From this relation of the shearing angles in the two types of rock it is evident that the joints in the more brittle beds are due to the normal component of the stress acting on the beds. They

[1] *Sixteenth Ann. Rept. U.S. Geol. Survey,* Part I (1896), p. 652. See also C. K. Leith, "Rock Cleavage," *U.S. Geol. Survey, Bull. 239* (1905), p. 123, Fig. 37.

must, therefore, have been developed essentially through irrotational strain.

The differential movement between adjoining beds required by the structural relations indicated in the text, was obviously largely limited to slippage parallel to the bedding within the thin slaty layers. It is important to note, however, that the direction of movement here, as in several cases referred to in the first part of this paper, has influenced the number and nature of the two opposed diagonal joint systems in the brittle beds. Those inclined in the direction of drag produced by the differential movement are more numerous, closed, and slickensided, while the opposite set is represented by few and gaping joints.

The closely spaced joints in the slaty beds, on the other hand, may partly be due to true rotational strain.

Essentially the same considerations apply to the case discussed in detail by Steidtmann[1] and by Leith in his *Structural Geology*. Here the joint system opposed to the "drag" action of the differential movement between the beds is represented by but a few "open gashes or tension joints." But their presence is sufficient to indicate that in the center of the quartzite beds the effect of rotational strain has been entirely subordinate to that of the irrotational strain produced by the component normal to the bedding planes.

The writer has the suspicion that this will be found to be true most generally in bedded rocks, in which differential movement between the beds is largely accomplished by slippage along bedding planes chiefly within layers of soft rock acting as lubricants.

V. PLANES OF SHEARING IN SHALES

The absolute and relative values of the crushing and tensile strengths of a rock play a fundamental roll in the deformation of rocks along the face of natural and artificial excavations.[2]

When the depth of an excavation has reached the point at which the vertical component of the stress set up by the removal of

[1] *Jour. Geol.*, Vol. XVIII (1910), p. 261; Fig. 1.

[2] See the excellent analysis of the factors involved in D. F. MacDonald, "Excavation Deformations," *Congrès géologique international, Compte rendu de la XIIe session* (Canada, 1913), pp. 779–92.

FIG. 16.—Fold and thrust fault in Eden shales, West Fork Creek, Cincinnati, Ohio

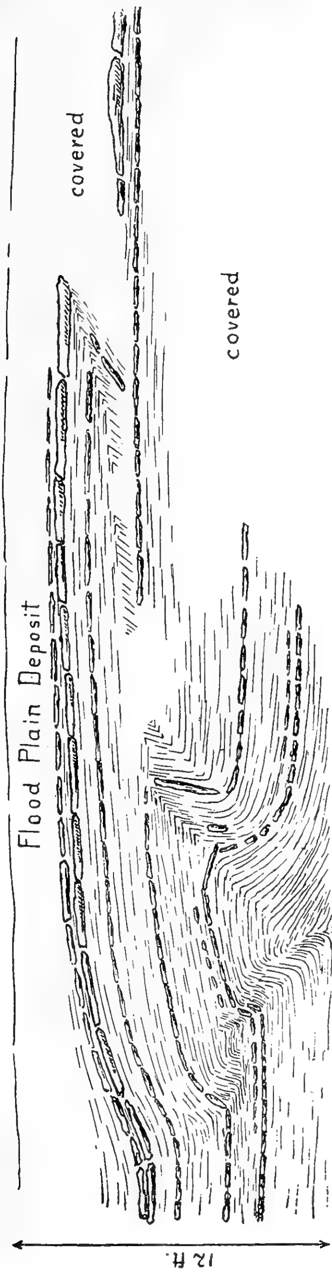

FIG. 17.—Fold and thrust fault in Eden shales, West Fork Creek, Cincinnati, Ohio

material at the toe of the new steep slope exceeds the elastic limit, rupture along planes of shearing will take place sooner or later in brittle materials. Ductile substances, on the other hand, such as

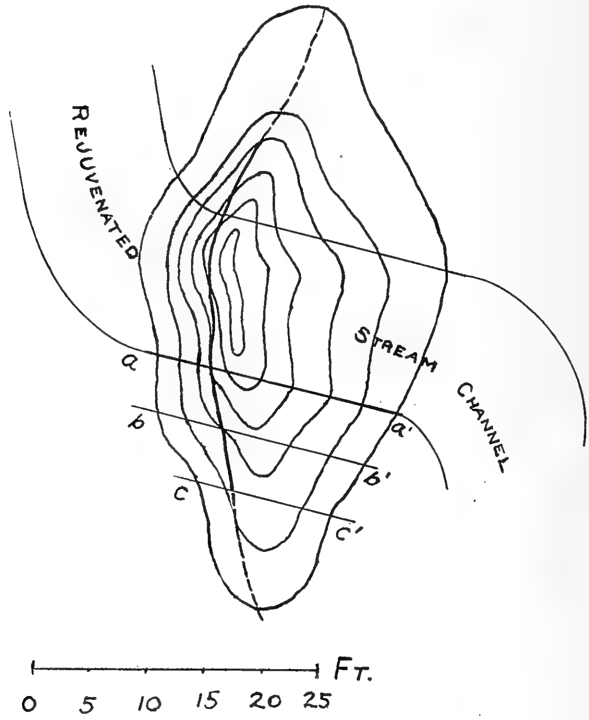

FIG. 18.—Diagrammatic contour map of the surface of a bed of shale buckled and cut by an inclined joint plane. Superimposed on it is the outline of a new stream bed cut into this structure after rejuvenation.

clays, will flow, causing the lower part of the steep slope to bulge out and the bottom of the excavation to buckle up.[1]

Van Horn has given a detailed description of this buckling at the base of a rock slide.[2] Small and entirely local anticlines which

[1] D. F. MacDonald, "Excavation Deformations," *Congrès géologique international, Compte rendu de la XII^e session* (Canada, 1913) p. 791, Fig. 3.

[2] Frank R. Van Horn, "Landslide Accompanied by Buckling, and Its Relation to Local Anticlinal Folds," *Bull. Geol. Soc. Amer.*, Vol. XX (1910), pp. 625–32.

quite obviously owe their origin to this process, are of common occurrence in the bottoms of ravines cut into clays or shales, and are often directly associated with landslip terraces.[1] Identical surficial anticlinal buckles[2] which have been observed under a cover of glacial drift without any connection with steep slopes or landslides, probably owe their origin to similar stress relations resulting from the cracking or other deformation of the Pleistocene ice cover.

In the vicinity of Cincinnati, wherever the most recent rejuvenation has cut into the bottom of ravines within the Eden shales, similar bucklings are quite common.[3] Frequently, however, these anticlines are not only overturned, but faulted, generally in the form of a miniature overthrust, such as shown in the Figures 16 and 17.

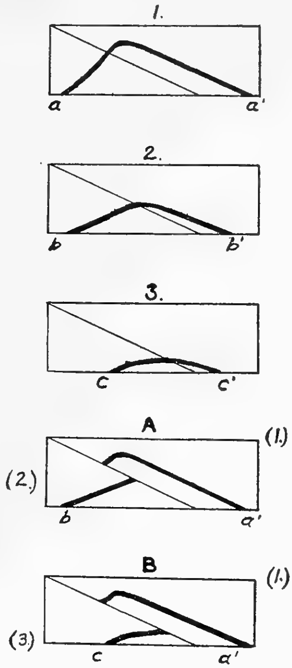

FIG. 19.—Sec. 1 shows the structure exposed on the south side of the stream channel shown in Fig. 18. Secs. 2 and 3 are drawn parallel to and 5 and 10 feet, respectively, south of, sec. 1. Sec. A shows the result of squeezing out the shale layers from underneath the joint plane in such a way that the lower portion of sec. 2 is pushed out so as to rest under the upper portion of sec. 1. In sec. B this process is carried farther, the lower portion of 3 pushed out so as to rest underneath the upper portion of 1.

In the light of the preceding discussion, it appears highly probable that these miniature "reversed faults" have nothing to do with horizontal compressive stresses. The shales are distinctly ductile, as is implied by the very existence of the anticlines due to flowage. They are, however, ductile only to a limited degree. After the strain has reached a certain limit, they rupture along planes of shearing. Here the obtuse angle formed by two

[1] E.g., T. C. Hopkins and W. M. Smallwood, "Discussion of the Origin of Some Anticlinal Folds near Meadville, Pennsylvania," *Bull. Syracuse University*, Ser. IV, No. 1, 18 (quoted from Van Horn, *loc. cit.*).

[2] For instance, G. K. Gilbert, "Dislocation at Thirtymile Point, New York," *Bull. Geol. Soc. Amer.*, Vol. X (1899), pp. 131–34; F. R. Van Horn, "Local Anticlines in the Chagrin Shales at Cleveland, Ohio," *ibid.*, Vol. XXI (1910), pp. 771–73.

[3] E. L. Braun, "The Cincinnatian Series and Its Brachiopods in the Vicinity of Cincinnati," *Jour. Cinc. Soc. Nat. Hist.*, Vol. XXII (1916), No. 1.

symmetrical joint planes points to the action of compressive stress in a perpendicular direction, or especially to the horizontal tensile

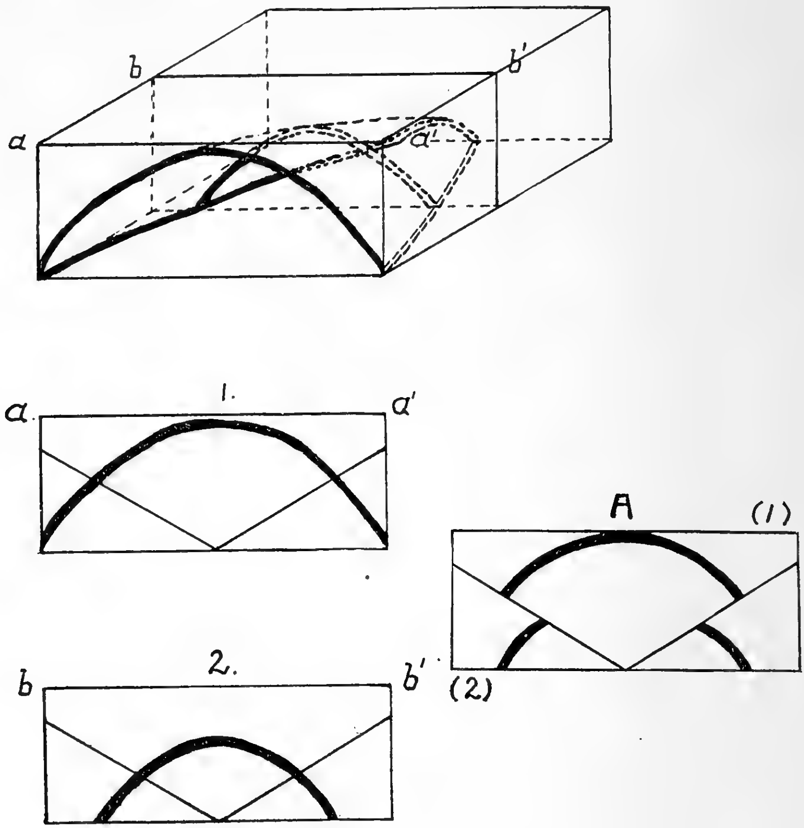

FIG. 20.—Block diagram representing the pitching end of an anticline.
Sec. a–a′ = sec. 1 below.
Sec. b–b′ = sec. 2 below.

These sections show, in addition, the position of two inclined joint planes striking parallel to the axis of the fold. Sec. A shows the result of squeezing out the shale layers from underneath the joint plane in such a way that the lower portion of sec. 2 is pushed out so as to rest under the upper portion of sec. 1.

stress which results when the upper layers are forced up by the pressure of the flowing layers of shale underneath.

Whenever a stream removes a portion of such an anticline, as indicated on the map sketch in Figure 18, the tendency exists to

squeeze the dipping beds up from underneath the joint plane, out into the channel.

Sections A and B (Fig. 19) show the result which is brought about entirely by a horizontal flowage of the beds beneath the joint plane in a direction perpendicular to the plane of the paper, and not to any horizontal compressive stresses acting within the plane of the paper.

The joint nature of these "fault" planes is brought out beautifully by the exposure shown in Figure 17, which for years has been well exposed in Westfork Creek at Cincinnati, just above the schoolhouse.

There can be little doubt that essentially the same interpretation applies to other similar occurrences.[1]

VI. HORIZONTAL COMPRESSIVE STRAINS IN GRANITE

Compressive strains of considerable magnitude, essentially in a horizontal direction, exist at widely separated localities in all states of New England, if not throughout the whole region. In the quarries, the strains find expression in various ways. Vertical drill holes are flattened to an elliptical cross-section,[2] the cores between contiguous borings are crushed, and cracks open up diagonally from the channels.[3] In the process of quarrying new fissures open up with a dull explosive noise,[4] or new sheetlike partings form[5] or old sheets buckle up.[6]

With a detailed knowledge of most of the excellent exposures of sheeted granite in New England at his command, Dale has come to the conclusion that shrinkage in cooling, or changes of temperature, or other forms of weathering have played, at best, only a secondary rôle in the production of the sheet structure; that the

[1] A. H. Purdue, "Illustrated Note on a Miniature Overthrust Fault and Anticline," *Jour. Geol.*, Vol. IX (1901), pp. 341–42; C. E. Decker, unpublished manuscript, see Fig. 17, p. 39, *Jour. Geol.*, Vol. XXVI (1918).

[2] T. N. Dale, "The Granites of Vermont," *U.S. Geol. Survey, Bull. 404*, p. 18, Fig. 2; *Bull. 354*, p. 34.

[3] *Ibid., Bull. 354*, pp. 96 and 126; *Bull. 313*, pp. 12 and 142.

[4] *Ibid., Bull. 313*, pp. 34 and 142.

[5] *Ibid., Bull. 404*, pp. 97 and 107.

[6] *Ibid., Bull. 313*, Pl. VII, A.

main factor has been horizontal compressive stress such as now finds expression in the region.[1]

This is not the place nor the time to enter into a discussion of this complex problem. The final word has not yet been spoken.

If we assume, for the time being, a diastrophic origin of the observed strains, we can point to three observations favoring such a view.

1. For at least one locality, the famous Quincy quarries south of Boston, Dale records the observation that "this strain in some quarries appears to increase with their depth."[2]

2. At Fletcher Quarry, on Robeson Mountain, in Washington County, Vermont, Dale observed what he called "double-sheet" structure. Here, instead of the usual single set of sheets, two such sets, intersecting at an angle of about 42° are exposed. If we analyze their position according to the method described in the first part of this paper we find that the bisectrix of the acute angle, that is, the direction of the greatest principal (compressive) stress, trends N. 60 W.–S. 60 E. and that it differs but slightly from the horizontal, being slightly directed downward toward the southeast. The least (tensile) principal stress, on the other hand, is practically directed upward, in the direction of easiest relief from the horizontal pressure.[3]

Fortunately, there is, at the same locality, "a marked northeast–southwest compressive strain in the upper part of the quarry, raising the sheets and even forming new sheet partings." The direction of this strain is essentially that which would result from the compressive stress, acting from the northwest, inferred from the "double-sheets." While this may, of course, be a mere coincidence, it certainly is suggestive of a causal connection.

None of the other quite numerous cases in which Dale records the direction of compressive strain, together with data concerning the position of the sheet structure, can be used to test this matter

[1] T. N. Dale, "The Granites of Connecticut," *U.S. Geol. Survey, Bull. 484* (1911), pp. 29–36.

[2] *U.S. Geol. Survey, Bull. 354*, p. 96.

[3] If we had reason to believe that the granite had been in a ductile state when these planes of parting were formed, the stresses would have to be reversed. But all observations seem to speak against this possibility.

further. For with only one of the two intersecting planes of shearing given, we cannot even guess at the position of the principal stresses.[1]

3. The New England region is that part of the United States of which we know most definitely and quantitatively that it has undergone considerable deformation in post-glacial time. A glance at Fairchild's map of "Isobases of Post-glacial Uplift"[2] shows that if we can assume the earth's surface to the south and southeast to be relatively at rest and the domelike upheaval with its center halfway between Quebec and James Bay to be rising independently, there should exist a compressive stress in the general direction northwest–southeast throughout the surface of New England. It may, of course, be mere coincidence that this is the same direction which we found required to produce the "double-sheet" structure on Robeson Mountain. But, combined, these three observations are distinctly favorable to the view that the larger part of the remarkable sheet structure of the granites of New England is due to compressive stresses arising from the larger earth movements indicated by Fairchild's isobasic maps.

The objection may be raised that in quarries in other parts of New England strains in different directions have been observed.

Great variations in local conditions of stress are not surprising in view of two important facts. The direction of relief in these cases is normal to the surface. Where the surface is far from horizontal the position of one of the two principal stresses must vary from place to place, and with it, the position of the resulting plains of shearing.

Furthermore, observation shows that the granites, which exhibit the sheet structure, are divided by various joint systems, some unquestionably of earlier origin. It is, therefore, not an unbroken sheet of granite that is being deformed in this region, but a mosaic of large and small blocks. The stress conditions, resulting

[1] If, for instance, in the case illustrated in Fletcher Quarry only the planes striking N. 20 W. had been given, we would probably not have considered them to be in harmony with a compressive stress acting from the northwest.

[2] H. L. Fairchild, "Post-glacial Uplift of Northeastern America," *Bull. Geol. Soc. Amer.*, Vol. XXIX (1918), p. 202.

from such a complex arrangement, must be expected to vary considerably from place to place, while corresponding only in the aggregate to the simple fundamental stress relation.[1]

Whatever may be the true merits of this interpretation of the "double-sheet" structure, it serves well to bring out the fact that horizontal compressive stresses will produce planes of shearing at low angles only in highly brittle substances close to the surface, where the least principal stress is directed upward, in the direction of easiest relief.

VII. LOW-ANGLE FAULTING

This leads us finally to the important question of low-angle faulting which has recently been given a most valuable discussion by R. T. Chamberlin and W. Z. Miller.[2]

The heart of the problem is nowhere clearer exposed than in the northwest Highlands of Scotland.[3] A series of sediments of diverse nature, piled up, shingle-like, by thrust faults inclined on the average about 45°, has been cut across by several practically horizontal planes of fracture, along which the individual slices have been moved in the general direction of thrust for many miles.

The strongly inclined faults clearly are planes of pure (irrotational) shear produced by the horizontal compressive stress. The

[1] For a similar reason, the movement on individual shearing planes, into which a block has been broken under one-sided pressure, may be most diverse, one block being pushed out in one direction, the other in another. This explains the small success that has attended the attempts made by Dr. Salomon (1911) and some of his students (Lind, 1910, Dinu, 1912, Spitz, 1913, Engstler, 1913, Seitz, 1917) to determine the nature of the stress involved in the formation of joint systems from the direction of movement recorded on the slickensides. (For complete references to these most careful studies, see O. Seitz., *Über die Tectonik der Luganer Alpen,* Inaug. Diss., Heidelberg, 1917).

The same criticism applies equally to the attempts made to infer from local observations of movements during an earthquake the nature of the major stresses which have resulted in the formation of great fault lines. The cause of such movements along existing fault lines may be fundamentally different from that which caused the fracture itself.

[2] R. T. Chamberlin and W. Z. Miller, "Low-Angle Faulting," *Jour. Geol.*, Vol. XXVI (1918), pp. 1–44.

[3] B. N. Peach, John Horne, W. Gunn, C. T. Clough, and L. W. Hinxman, "The Geological Structure of the North-West Highlands of Scotland," *Mem. Geol. Survey of Great Britain*, 1907.

question is, Can the essentially horizontal planes of fracture have been produced under similar conditions of strain or do they represent the result of rotational strain?

To prove the first assumption, we must find a cause for the great change in inclination of the shearing planes from rather high angles to a horizontal position. Cadell's experiments and their own researches led Chamberlin and Miller to the suggestion that the additional vertical pressure resulting from the piling up of numerous thrust-blocks may tend to reduce the angle of shearing.

In the light of the preceding discussion it appears that an additional load would rather have the opposite effect. It would be equivalent to increasing the hoop pressure in Kármân's experiments. It would render the rock less brittle and therewith the angle of shearing larger. Since the compressive stress, in this case, acts in a horizontal direction this would mean planes of shearing inclined at a steeper angle than before.

Any given rock shows the smallest angle of shearing at the surface. To reduce this angle still more, an active tensile stress would have to be applied in a vertical direction, which is, of course, out of question.

In a general way, then, we must picture to ourselves the planes of pure shear produced in the earth's crust by a horizontal compressive stress as being least inclined near the surface and becoming progressively steeper at lower levels so as to approach finally the more or less vertical position of the planes of flowage existing at considerable depths. Conceptions such as are expressed, for instance, in Ulrich's diagrams to illustrate the "inland migration of belts of folding in Southeastern North America" in his "Revision of the Paleozoic Systems"[1] require corresponding modification.

For an understanding of the mechanics of the large horizontal overthrusts, however, we can but turn to the only alternative, which was clearly brought out by Chamberlin and Miller, that, in contrast to the other type, they are essentially the result of rotational strains of the kind which Becker has termed "scission," which occur "when a bar or plate is shorn by a pair of shears, or

[1] *Bull. Geol. Soc. Amer.*, Vol. XXII (1911), p. 441.

when a rivet yields perpendicularly to its axis, say, in a bursting boiler."[1]

A clear and simple illustration of the form in which this method of fracturing finds its most important expression in the earth's structure, is offered by Figure 21, which represents the section of one of the folds of the Jura Mountains at the boundary of France and Switzerland, as brought out by Buxtorf's detailed investigations.[2] The brittle layer of the Rauracien, which stood up vertically in the process of folding, comparable to the rivet on a boiler plate mentioned above, gave way to the pressure of the swelling plastic Oxford clays, counteracted below by the level part of the north limb of the anticline.

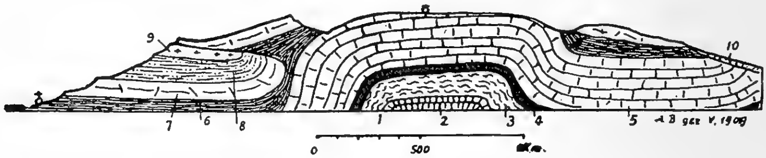

FIG. 21—Cross-section of a fold of the Jura Mountains at the border of France and Switzerland (at the "Clos du Doubs"). 1-3, Triassic, 4. Liassic, 5. Dogger, 6. Oxford clays, 7. Rauracian, Sequanian, 9. Kimmeridgian, 10. Landslide. (A. Buxtorf, 1909.)

The writer is inclined to believe, following the lead of Dr. Buxtorf,[3] that in the case of most, if not all "low-angle faults," or as Suess[4] called them, "listric planes," *the rotational stress resulted from the flowage of materials inside the fold*, whether it be, on a smaller scale, confined to individual normally plastic layers, or, on the largest scale, to the forced flowage of the rocks forming the cores of rising mountains.

This, however, leads us beyond the scope of this paper. The necessary setting for a broader discussion of this important problem will be given in a paper now in preparation.[5]

[1] G. F. Becker, "Finite Homogeneous Strain, Flow and Rupture of Rocks," *Bull. Geol. Soc.*, Vol. IV (1893), p. 24.

[2] A. Buxtorf, "Über den Gebirgsbau des Clos du Doubs und der Vellerat-Kette im Berner Jura," *Ber. Vers. Oberrhein. Geol. Ver.* (1909), p. 76, Fig. 1.

[3] *Ibid.*, p. 86.

[4] E. Suess, *The Face of the Earth* (transl. by Sollas, H. B.), Vol. IV (1909), p. 536.

[5] Parts of which were presented before the Geological Society of America at the Chicago meeting 1920 under the title "The Probable Cause of the Localization of the Major Geosynclines."

FEATURES OF A BODY OF ANORTHOSITE-GABBRO IN NORTHERN NEW YORK[1]

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GENERAL RELATIONS OF THE ANORTHOSITE-GABBRO

The body of rock considered in this paper varies from true anorthosite, through anorthosite-gabbro, to true gabbro. It is four and one-third miles long, with a maximum width of one mile. It lies a little west of the central part of the Russell quadrangle in St. Lawrence County, New York (Fig. 1).

As far as could be determined, the rock immediately surrounding the gabbro is granite which is generally pink in color and carries only moderate amounts of dark minerals. The granite varies from medium grained to coarse grained, and from scarcely foliated to highly foliated. That the granite is younger than the anorthosite-gabbro is proved by the presence of both granite and granite-pegmatite dikes in the gabbro, and by the more or less intimate injection of portions of the borders of the gabbro by the granite. The best and most instructive exposures occur within the northern one-third of the area.

About twenty-five years ago Professor C. H. Smyth published a short paper[2] which is largely a microscopical petrographic description of a ledge of gabbro probably lying within the area of anorthosite-gabbro. It is the present purpose to rather fully discuss the megascopic and microscopic features, field relations, and origin of the remarkable lot of rocks which constitute facies of the body of anorthosite-gabbro and directly associated rocks covering several square miles.

VARIATIONS IN COMPOSITION

The whole mass of the rock of the area is called anorthosite-gabbro because it shows every possible gradation from true gabbro

[1] Published by permission of the state geologist of New York.

[2] C. H. Smyth, *Amer. Jour. Sci.*, 4th series, Vol. I (1896), pp. 273–81.

(light gray to dark gray) to almost pure plagioclase anorthosite (light gray to nearly white). In the northern portion of the area the light-gray anorthositic facies are most conspicuously developed, but they occur locally in all parts of the area.

In thin section under the microscope the feldspars of the gray to nearly black facies contain myriads of dustlike inclusions,

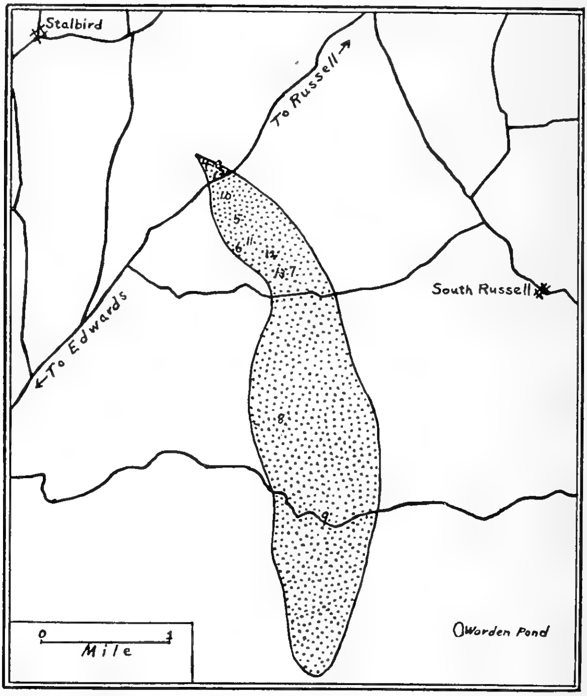

FIG. 1.—Sketch map of part of central St. Lawrence County, New York, showing the location of the body of anorthosite-gabbro. Numbers indicate localities referred to in the text.

probably ilmenite. No. 42 of Table I (see below) represents a thin section of a specimen of the typical very common dark-gray, medium-grained to moderately coarse-grained, non-foliated gabbro with ophitic texture from locality 9 (see Fig. 1). The labradorites, which range in length up to one-half an inch, are generally euhedral. All the pyroxenes (probably diallage) have narrow rims of hornblende around them, but some hornblende occurs either in small

clusters or as scattering grains. There is some very local granulation, especially of the feldspars.

No. 39 of the table represents a thin section of a light greenish-gray, medium-grained, moderately ophitic gabbro from locality 8. The andesine is much altered, mostly to scapolite. The pyroxene (probably diallage) is pale green to colorless. Tiny black inclusions in it are common. Most of the hornblende forms distinct granular rims around the pyroxenes, but a good deal of it is elsewhere interlocked with the pyroxenes. There is some granulation of the feldspars in local zones only.

No. 51 of the table is from a specimen of the very common dark-gray, non-foliated, medium-grained to fine-grained gabbro with ophitic texture from locality 7.

At locality 6 the main body of the rock is a medium-grained to moderately coarse-grained gabbro with a crude ophitic texture. The feldspar laths range up to three-fourths of an inch in length. No. 47 of the table shows the minerals contained in a thin section. Common green hornblende occurs mostly as narrow rims around both pyroxene and magnetite. Small inclusions of hornblende occur in the plagioclase. Granulation is lacking. Lying within the rock just described there is a zone of distinctly porphyritic, non-foliated gabbro about a rod wide, without sharp contacts. A hand specimen shows scattering, irregularly arranged, broad, dark laths of plagioclase up to one-half of an inch in length imbedded in a fine-grained dark-gray groundmass. No. 46 of the table shows the minerals observed in a thin section. Much of the feldspar has been altered, chiefly to scapolite. There is apparently no granulation, but considerable masses of fine granular hornblende, pyroxene, biotite, and plagioclase fill large spaces between, and cracks in, the feldspar phenocrysts. A good many grains of biotite, hornblende, and pyroxene are also irregularly scattered through the feldspars.

In a ledge at locality 1 most of the rock is a coarse-grained hornblende-rich facies of the gabbro. This rock is light gray, due to whiteness of the feldspars, which range in length from the merest fraction of an inch to several inches. Most of the feldspars are distinctly euhedral. A coarse ophitic texture is usually well

exhibited over the weathered surface. No. 57 of the table represents a thin section. The large euhedral crystals of plagioclase (chiefly labradorite) are generally fringed with rather clear granular scapolite, while their interiors are considerably decomposed. On one side this rock is moderately foliated, the feldspars being drawn out into crude parallelism. This foliated portion is not sharply separated from the rest of the rock. Within the foliated portion, and not sharply separated from it, there is a tongue or dikelike mass of gabbro very rich in dark minerals, chiefly hornblende.

TABLE I

Minerals by Volume Percentages in Thin Sections of the Anorthosite-Gabbro and Closely Associated Rocks

Slide Number	Localities Shown on Sketch Map	Labradorite	Andesine	Oligoclase to Labradorite	Andesine to Labradorite	Scapolite	Augite or Diallage	Hornblende	Biotite	Quartz	Chlorite	Magnetite	Pyrite	Hematite	Zircon	Apatite	Titanite	Epidote
39	8	....	66	....	....	....	25	6	½	....	Little	2	....	....	Little	....	....	½
42	9	74	....	....	....	....	16	5	2	½	....	2	½	....	....	....	....	....
46	6	....	....	....	82	....	7	7	4	....	....	....	....	....	....	....	....	....
47	6	....	....	....	76	....	15	3	1½	....	....	4	....	....	....	½	....	....
51	7	....	....	50	....	....	24	10	6	....	....	8	½	1½	....	....	....	....
57	1	....	....	64	....	10	....	25	....	....	....	1	....	....	....	....	....	....
56	1	....	....	76	....	....	10	12½	....	....	....	½	....	....	1	....	....	....
58	1	....	....	....	59	....	14	23	....	....	....	Little	....	....	4	....	....	....
59	1	46	....	....	....	....	26	26	....	....	....	1	....	....	1	....	....	....
52	7	....	....	....	....	39	....	43	....	13	....	4½	½	....	....	....	....	....
49	5	....	....	64	....	35	....	½	....	....	....	....	....	....	Little	Little	¼	....

In the coarse-grained gabbro at locality 1, less than a rod from the rocks above described, there is a highly foliated zone or belt of very variable rocks about a yard wide. Much of this rock is fine grained, dark gray, highly foliated, and uniform in appearance, a thin section being represented by No. 51 of the table. There appears to be no true granulation. Another facies of this zone is a highly foliated, medium-grained, light-gray gabbro which in thin section is represented by No. 56 of the table. Still another facies consists of fine-grained, highly foliated, alternating bands of practically pure white plagioclase and gray gabbro, these bands varying in width from one-tenth of an inch to an inch or two. No. 58 of the table gives the minerals contained in a thin section of this third facies.

An important feature, excluding the dikes below described, is the close association of facies very different in composition. In some cases such facies are asymmetrically arranged, but more commonly they are in the form of more or less well-defined belts of zones which in some cases show fairly sharp boundaries, though they mostly grade into the adjacent facies. Such belts or zones commonly vary from a fraction of an inch to some yards or rods wide. The general effect is more or less well-defined local banded structures within the body of anorthosite-gabbro, especially in its northern portion. (See Fig. 2.)

VARIATIONS IN FOLIATION

Many portions of the anorthosite-gabbro, especially toward the north, exhibit more or less well-defined foliation. The greatest bulk of the rock is, however, practically devoid of such a structure. The non-foliated rock is mostly true gabbro, varying from light gray to very dark gray in color, medium grained to moderately coarse grained, generally with an ophitic texture. Some of the light-gray anorthosite-gabbro and anorthosite facies are also non-foliated. In many parts of the whole area there are gradations from facies which exhibit little foliation to others which are very highly foliated. As in the case of the variations in composition, so here the variations in degree of foliation generally occur as more or less well-defined belts or zones. Such belts commonly range in width from a few inches to some yards or rods. Highly foliated, moderately foliated, and non-foliated belts may lie adjacent to each other, in some cases with fairly sharp contacts, and in many cases with gradations from one into another. The variations in foliation are best exhibited in the northern portion of the area. In certain cases dark facies of the gabbro are so highly foliated that they look like hornblende schists. In other cases the rocks look like typical basic gneisses with alternating light and dark bands. Most of the anorthosite or gabbro shows moderate foliation due to a crudely parallel arrangement of the minerals, especially the dark minerals, without producing a banded effect. Figure 2 gives a fair idea of the variations in composition and foliation in a single small ledge.

An important feature of the foliation is the notable variation in strike within the body of anorthosite-gabbro. Ranges in strike

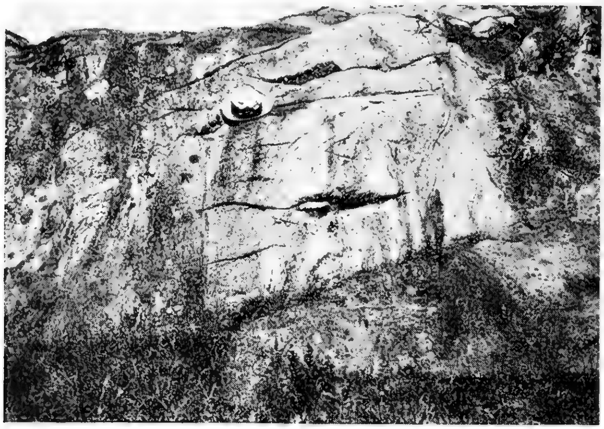

FIG. 2.—An exposure of anorthosite-gabbro showing notable local variations in composition and foliation. Just north of the northern road across the area.

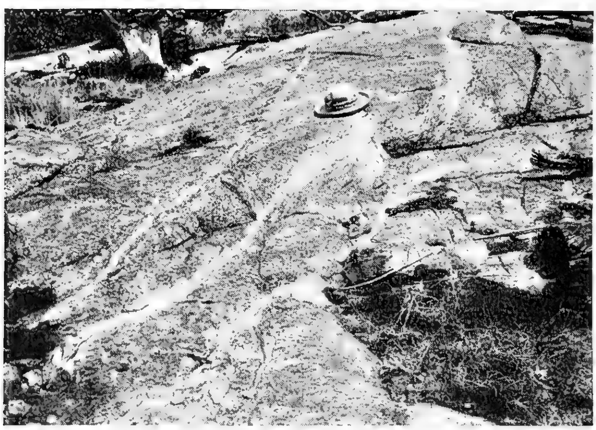

FIG. 3.—White dikes rich in plagioclase and scapolite cutting anorthosite-gabbro, one-third of a mile south-southeast of the northern road.

from northeast to northwest have been observed. Notable variations also occur within relatively short distances, even within a few yards or rods.

VARIATIONS IN TEXTURE

As already stated, most of the rock, except toward the north, is a true gabbro varying from gray to very dark gray, the latter predominating. Such gabbro is medium grained to coarse grained, and it commonly possesses a fair to excellent ophitic texture. In the northern portion of the area most of the rock is a gray to light-gray, medium-grained to moderately coarse-grained anorthosite-gabbro varying to anorthosite which does not so commonly possess an ophitic texture. Local zones of light- to dark-gray anorthosite-gabbro and gabbro throughout the area are fine grained to medium grained and generally more or less highly foliated. In some local portions of the general mass almost perfectly euhedral crystals of plagioclase up to an inch long are set in a fine, granular groundmass of hornblende and monoclinic pyroxene.

The great bulk of the dark-gray gabbro shows little or no granulation. The very light-gray anorthositic facies are commonly considerably granulated. Much of the white feldspar of the light-gray anorthosite-gabbro, especially in the northern portion of the area, shows more or less granulation. Non-foliated facies of the light-gray anorthositic gabbro exhibit little or no granulation of the dark minerals, but not uncommonly in the foliated facies the dark minerals as well as the feldspar are granulated.

CAUSES OF THE VARIATIONS IN COMPOSITION, STRUCTURE, AND TEXTURE

It is believed that most, if not all, of the foregoing variations in composition, structure, and texture are primary features, in other words, that they developed under conditions of moderate pressure before or during the final consolidation of the anorthosite-gabbro magma. The main principles involved in the development of such features have been set forth by the writer in several recent papers[1] dealing with the origin of the foliation and banding of both the anorthosite and syenite-granite series of the Adirondack region.

[1] W. J. Miller, *Jour. Geol.*, Vol. XXIV (1916), pp. 600–619; *Science*, N.S., Vol. XLVIII (1918), pp. 560–63; *Bull. Geol. Soc. Amer.*, Vol. XXIX (1918), pp. 399–462.

A brief statement of the writer's explanation of the origin of the variations in composition, structure, and texture of the anorthosite-gabbro body will now be given.

1. It is believed that the intrusion of the anorthosite-gabbro magma was a complicated process of considerable duration. Much of the magma slowly solidified as a normal, moderately coarse-grained to medium-grained gabbro with an ophitic texture. Magmatic currents were practically absent from those portions.

2. In portions of the magma there were currents or movements sufficiently strong to cause the minerals, especially the earlier crystallized ones, to be drawn out into more or less perfect parallelism. This accounts for the gneissoid structure or foliation of considerable portions of the gabbro and anorthosite-gabbro, especially the medium-grained and fine-grained portions.

3. The more or less highly foliated structures of the many localized zones or belts of medium-grained to fine-grained rocks resulted from particularly active currents in those zones.

4. The more or less sharp alternations or variations in degree of foliation, and the notably varying strikes of the foliation throughout the area, are best explained as a result of locally developed differential magmatic movements or currents under conditions of not more than very moderate pressure. It seems impossible to account for such structures on the basis of pressure applied upon the body of anorthosite-gabbro after its final consolidation, first, because such pressure could not have caused notable foliation of certain narrow belts while adjacent masses were unaffected, especially in view of the fact that the highly foliated zones are not characterized by shearing or crushing (granulation), and, second, because any such pressure sufficiently strong to produce foliation must have been exerted in a single general direction with about equal force, and hence the sharp variations in strike of the foliation are left unaccounted for.

5. The variations in composition are the result of differentiation of the anorthosite-gabbro magma before or during its consolidation.

6. Those foliated zones or belts which show sharp variations in composition in layers ranging from nearly pure plagioclase to gabbro rich in ferro-magnesian minerals are the result either of

differentiation into such layers, probably accompanied by differential movements, during the intrusion, or of local intrusions of heterogeneous portions of the magma in which the differentiation took place before intrusion along local zones.

7. Certain local dikelike facies extra rich in ferromagnesian minerals, which are not sharply separated from the main mass of the anorthosite-gabbro and send dikelets into it, are basic differentiates from a lower level forced upward into the more normal mass which had already more nearly approached final consolidation.

8. The granulation of the feldspars in those portions of the foliated lighter-gray gabbro and anorthosite-gabbro in which the ferromagnesian minerals show little or no granulation is believed to have resulted mainly from movements in the magma before final consolidation, that is during or after the crystallization of the feldspars, but while the ferromagnesian minerals were still more or less liquid. In other words, we are here dealing with a protoclastic texture. Such pressure could not have been applied after complete consolidation of the magma because, if so, the dark minerals would also be granulated. Where the ferromagnesian minerals also show granulation, this probably resulted from magmatic movements just prior to final solidification, or from pressure upon solidified portions of the mass exerted by locally intruding portions of still liquid rock. The possibility is conceded, however, of the development of some granulation of this type by external application of moderate pressure upon the whole body of anorthosite-gabbro, but the localization of such granulation renders such an explanation less likely.

DARK DIKES RICH IN HORNBLENDE AND SCAPOLITE

At locality 7 there is a fine display of rather remarkable dikes. The country rock is there mainly a medium-grained, dark-gray, rather typical, non-foliated gabbro with ophitic texture. No. 51 of the table gives the mineral content of a thin section.

Over an area of several acres the rock just described is cut most irregularly by a network of numerous dikes of dark-gray, highly foliated, fine-grained to medium-grained rocks consisting chiefly of scapolite and hornblende together with some quartz and

magnetite and a little pyrite. No. 52 of the table shows the simple but rather remarkable mineral content of a thin section of one of these dikes. The scapolite is clear and fresh in the form of irregular grains. The hornblende is the common variety with pleochroism ranging from dark green to greenish yellow. Nearly all the minerals are distinctly elongated parallel to the foliation. There seems to be no granulation. These dikes have more or less indefinite boundaries against the normal gabbro into which they fray out most irregularly. They seldom exceed a foot in width. The highly foliated structure of dikes arranged like these in normal gabbro cannot have resulted from pressure brought to bear upon the whole mass of rock after it had all become consolidated. It is evident that we are here dealing with a remarkable case of primary foliation developed to a high degree in small dikes before their final consolidation. The lack of sharp contacts points to the intrusion of the dike material into the gabbro while the latter was still hot, though nearly or quite consolidated.

In the light-gray anorthosite-gabbro of the northern portion of the area, some dark, hornblende-rich, well-foliated dikelike bands occur. In some cases the contacts are fairly sharp, and in others they are not. Most of these bands quite certainly represent intrusions into the still hot but nearly or quite solidified country rock, and their foliation is of primary origin. None of these bands show a network arrangement. Whether or not they, too, are rich in scapolite is not known.

WHITE DIKES RICH IN PLAGIOCLASE AND SCAPOLITE

At a number of localities white dikes were observed to cut the anorthosite-gabbro and gabbro. A good display of such dikes may be seen in a ledge at locality 5 (see Fig. 3). The main rock of the ledge is gray, medium-grained, well-foliated gabbro varying to anorthosite-gabbro. Several white dikes ranging in width from one to several inches cut across the foliation of the gabbro at high angles. These dikes, which are roughly parallel and traceable for a rod or more, contain a maximum of not more than a few per cent of dark minerals. They are fine grained with some scattering feldspars up to one-half an inch in length. Contacts against the

country rock are not very sharp. No. 49 of the table gives the mineral content of a thin section of the dike rock. Little or no granulation is shown. The scapolite, which is very clear and fresh, occurs mostly in the form of irregular grains as inclusions in the feldspar. Since the scapolite grains are variously oriented in the feldspar and their contacts are sharp, it is evident not only that the scapolite crystallized before the plagioclase, but also that it is not merely a decomposition product of the feldspar. Either the scapolite is of primary origin, or it somehow represents recrystallized material.

On a little hill at locality 10 several similar dikes were observed. Some of these show a moderate degree of foliation. Others occur here and there in the anorthosite-gabbro at localities 6, 7, 11, 12, and 13. Still others fill many of the cracks in the prominent zone of fractured and crushed anorthosite-gabbro described below.

On the one hand, the true dike character of these white rocks and their strike across the foliation of the country rock at high angles prove that they were intruded after the anorthosite-gabbro, which contains them, had almost or completely solidified. On the other hand, their lack of sharp contacts indicates that they were intruded while the anorthosite-gabbro was still hot enough to allow some merging of the dike materials with the walls. The dike material probably originated by differentiation in a still more or less liquid portion at a lower level and was then forced into fissures in a higher, cooler portion of the anorthosite-gabbro body. The dikes are, according to this view, satellitic dikes of the anorthosite-gabbro.

HORNBLENDITE DIKES

These are remarkable dikes consisting of pure hornblende. Altogether in the northern one-third of the area of anorthosite-gabbro many dozens of the hornblendite dikes were observed. They are generally grouped in certain portions of the anorthosite-gabbro where it has been fractured. At locality 10 many cracks, commonly in the form of a network in the light-gray, moderately coarse-grained, little foliated gabbro, are filled with small dikes of moderately coarse-grained pure hornblende. Contacts of the

dikes are not sharp. At locality 5, close to the white dikes shown in Figure 3, the anorthosite-gabbro contains many roughly parallel

FIG. 4.—Dikes of hornblendite filling curving cracks in anorthosite-gabbro, one-third of a mile south-southeast of the northern road.

FIG. 5.—Ledge of granite containing curving cracks filled with pyroxenite dikes on the left side, and nearly vertical cracks filled with pyroxene-rich pegmatite dikes on the right side. Close to the very northern end of the anorthosite-gabbro area.

curving cracks filled with small dikes of pure hornblendite (see Fig. 4). Still farther south, especially at localities 7, 11, 12, and 13, more cracks are filled with hornblendite dikes.

The hornblendite dikes are commonly from one-half an inch to one inch wide, and they are seldom traceable individually for more than a few yards or rods. They cut the anorthosite-gabbro in all directions, not uncommonly at high angles across its foliation. Contacts against the country rock are always sharp. The dike material is practically pure, common hornblende which, in the hand specimen, is greenish black where fresh, and dull green where weathered. The dikes are rather uniformly moderately coarse grained, the crystals usually varying in length from one-fifth of an inch to one inch. There is no suggestion of finer grain along the borders. Granulation is absent.

Many of the dikes occupy curving cracks in the anorthosite-gabbro as described below. In some such cases hornblendite dikes lie in rather sharp contact against dikes of nearly pure plagioclase anorthosite. On the southern face of the little hill at locality 1c, several hornblendite dikes cut sharply across some branching, white, plagioclase-scapolite dikes, the latter in some cases being faulted where the hornblendite dikes cross.

Both the hornblendite dikes and the white dikes rich in plagioclase and scapolite are quite certainly late differentiates of the cooling anorthosite-gabbro magma at some depth below the present rock surface. They are in a real sense complementary dikes, the differentiated white-dike material having been forced upward into the nearly or quite consolidated, but still hot, anorthosite-gabbro, while the differentiated hornblendite was later forced upward into cracks in the relatively much cooler body of anorthosite-gabbro. This is in harmony with the lack of sharp contacts of the white dikes, and the sharpness of contacts of the hornblendite dikes.

In the discussion of this paper at the Boston meeting of the Geological Society, N. L. Bowen suggested that the fissures filled with hornblende might better be called veins instead of dikes. But, because they show no vein structure and are in almost every way like true dikes, the writer considers them to be dikes, probably in the category of pegmatites because of their coarseness of grain and lack of chilled borders. It is accordingly reasonable to believe that the hornblende was either in solution in a pegmatite

residue of the anorthosite-gabbro magma, or it was taken into solution by magmatic solvents during their passage through the anorthosite-gabbro after it was nearly or wholly consolidated but still hot. That the general rock mass was still hot during the intrusion of the hornblendite dikes is proved by the fact that still later minor intrusions, as distinct facies of the anorthosite-gabbro, took place. In any case we are dealing with magmatic intrusive material (hornblendite) which should be classed as dike rather than vein material.

FRACTURED AND CRUSHED ZONES

A remarkable zone or belt of fractured, and in part crushed, anorthosite-gabbro occurs near the northern end of the area. It strikes about north 60° west and extends partly across the northern road near the granite contact at locality 2. This fractured zone maintains a fairly uniform width of about one rod, and it is very clearly exposed across a great ledge of bare rock for approximately two hundred feet on the north side of the road. South of the road it shows much less. The main body of rock of this zone varies from nearly white anorthosite to light-gray anorthosite-gabbro. This zone is characterized throughout its length by a remarkable system of curving, roughly parallel cracks. In a sense the cracks are festooned one within the other. This system of cracks extends across the zone of fracturing rather than parallel with it. Fig. 6 gives a fairly good idea of the structure, though the foreshortening in the picture is strong. The description immediately following pertains to this photographed locality twelve or fifteen rods north of the road. Between the cracks the distances generally range from one to six or eight inches. Most of the cracks (now filled) are from one-fourth of an inch to two inches wide, but they vary considerably even within a few inches. A single crack is seldom traceable for more than five or six feet. The cracks are filled chiefly with rather fine-grained, nearly pure-white, plagioclase-rich material. The walls of the cracks are mostly exceedingly irregular, with multitudes of sharp indentations, usually less than an inch deep, into which the feldspar has been filled. In many cases the plagioclase-rich filling sends off small tongues or dikelets, rarely over a foot long, ending in fairly sharp points in the anorthosite-gabbro. Some of the filled cracks are

distinctly lenslike, up to a foot long. That the plagioclase-rich material was forced into the cracks in the form of dikes seems

FIG. 6.—A portion of the remarkably fractured zone of anorthosite-gabbro with its dikes in the northern portion of the area. About fifteen rods north of the northern road.

FIG. 7.—A local highly crushed portion of the prominent zone of fracturing in the anorthosite-gabbro. A few rods north of the northern road.

almost certain. Contacts of the dikes against the anorthosite-gabbro are generally fairly sharp. On the north side of the photographed locality (see left side of Fig. 6) a good many very small

dikes of hornblendite occur within the dikes of white plagioclase-rich material. These dikes of hornblendite are exactly like the ones above described. Lying in the midst of the fractured zone at this locality and parallel to its strike, there is a twenty-inch wide band of dark, highly foliated gabbro rich in ferromagnesian minerals (see Fig. 6). This band of dark gabbro cuts squarely across the curving, dike-filled cracks by sharp contact on one side and by moderately sharp contact on the other side. In the midst of the dark band and parallel to it there is a dikelike strip of white plagioclase-rich anorthosite several inches wide, separated into blocks by multiple faulting. These faults extend across the dike-filled fractures of the anorthosite-gabbro on one side. The dark band of gabbro with its faulted white strip is well shown in Figure 6.

The history of the features which occur at the locality just described is probably about as follows. The main mass of the anorthosite-gabbro now at the surface was intruded and nearly or completely solidified. Then it was subjected to a torsional strain, probably due to a moderate pressure externally applied, whereby the system of parallel curving cracks was developed. The cracks were filled with plagioclase-rich dike material forced upward as a differentiate from a lower level of still more or less liquid gabbro. Still later came the intrusion of the hornblendite dikes into some of the same cracks, this material also having been produced as a differentiate at a lower level. Distinctly later, the twenty-inch-wide band of dark gabbro rich in ferromagnesian minerals was intruded, this material also having been derived from a portion of the general gabbro mass which was still in a more or less liquid condition at a lower level. Then the small white dike was intruded into the band of dark gabbro. Finally, after the whole complicated mass of gabbro and anorthosite now at the surface was solidified, the band of dark gabbro with its small white dike and the adjacent rock on one side were subjected to multiple faulting on a small scale.

A local portion of the fractured zone lying a few rods north of the road exhibits a high degree of crushing. Figure 7 gives a fair idea of this locality. The rock is mostly a very light-gray

to nearly white anorthosite relatively free from dark minerals. The ledge contains thousands of rounded to subangular pieces of medium to coarse granular plagioclase-rich material imbedded in a matrix of distinctly finer-grained groundmass. Remnants of the system of parallel curving cracks are clearly visible in parts of the ledge as, for example, just above the hammer in the picture (Fig. 7). Evidently here, as well as in that part of the ledge farther west, the system of crudely parallel curving cracks was developed by moderate pressure (probably a torsional strain) after the magma was nearly or completely solidified, but under continued pressure the rock was mostly shattered into small fragments. It seems likely that considerable nearly pure anorthosite magma then invaded the shattered mass and that the rounding off of the fragments was probably partly due to the mechanical crushing and partly to magmatic corrosion.

Certain other portions of the anorthosite-gabbro also show multiple fracturing but on lesser scales as, for example, at localities 5, 7, and 12.

PEGMATITE DIKE RICH IN PYROXENE

A pegmatite dike of special interest occurs in the anorthosite-gabbro at locality 3. It cuts across the foliation of the gabbro in the form of a lens with a length of twenty-five feet and a maximum width of three feet. It sends off one distinct branch about a yard long into the gabbro. This dike consists mostly of feldspar (chiefly oligoclase) and green monoclinic pyroxene, together with 5 to 8 per cent of quartz, all in anhedral crystals commonly from one to several inches long. Nearly all the pyroxene occurs in very irregular segregation masses up to several feet across. Some of the quartz is associated with plagioclase, and some occurs within the pyroxene segregation masses. This dike lies in sharp contact against the anorthosite-gabbro except within a few feet of each end.

This pegmatite is believed to be an offshoot from the adjacent granite, the magma or magmatic juices of which, on their way through the gabbro, dissolved or digested materials (chiefly ferromagnesian) from the gabbro, thus giving rise to the rather basic pyroxene-rich pegmatite. Because of the dissolving power of the

pegmatitic juices of magma, sharp contacts against the country rock would not be expected except where the magma had risen far enough to have lost its dissolving power. Further light is thrown upon the problem of the origin of this pegmatite rich in pyroxene in the discussion of the pyroxenite dikes (see below) which occur in the adjacent granite.

PYROXENITE DIKES IN THE ASSOCIATED GRANITE

Near the very northern end of the area of anorthosite-gabbro (locality 4), pyroxene-rich dikes cut a ledge of medium-grained, well-foliated, moderately hornblendic, pink granite close to its contact with the gabbro. Figure 5 shows the general relationships. On the left side curving parallel cracks in the granite are filled with dikes of pure to nearly pure, moderately coarse-grained, dark-green monoclinic pyroxene (apparently common augite). They contain only a few scattering crystals of orthoclase and quartz. None of these dikes is over an inch in width, and they all pinch out in the granite. Seven of them, up to two yards long, are perfectly arranged as a curved series one above the other in the face of the ledge. They show sharp contacts, and they cut across the foliation of the granite at high angles. On the right side (see Fig. 5) there are five or six coarser, crudely parallel, nearly vertical dikes, up to several inches each in width, in sharp contact with the granite. Crystals range up to two inches across. These dikes contain scattering crystals of orthoclase and quartz, but never more than 10 per cent. Between the two sets of dikes just described there is an irregular mass of very coarse pegmatite with crystals of quartz, orthoclase, and pyroxene up to six inches in length, but here the pyroxene is subordinate in amount. This mass shows only moderately sharp contacts against the granite.

It seems most probable that the pyroxene of both the pure pyroxene and pyroxene-rich pegmatite dikes was derived from the anorthosite-gabbro. Where the magma or juices passed through the gabbro, the pyroxene is believed to represent dissolved or digested and recrystallized ferromagnesian material.

It is important to bear in mind that the dikes of pure pyroxenite, as well as the dikes of hornblendite above described, are examples

of almost perfect mono-mineralic dikes. The pyroxenite dikes quite certainly, and the hornblendite dikes most probably, are of pegmatitic, magmatic origin, the magmatic solvent or solvents (probably largely water) having disappeared during the crystallization. In conclusion it is suggested that mono-mineralic dikes like the pyroxenite and hornblendite above described should not necessarily, as Bowen[1] reasons, be excluded from the category of intrusive rocks. Is it not true that all active magmas are more or less rich in mineralizers, and is it not further an open question as to whether any magma can remain active, that is, still possess an intrusive power, after its mineralizers have all disappeared? Pegmatitic magmas of relatively early origin in their parent-rock are probably such powerful agents of intrusion and even of injection because of their richness in mineralizers. Does not the argument that no rock consisting of but one mineral could ever have existed as an intrusive magma, because at least one mineral must act as a solvent and another as a solute, disregard the perfectly plausible possibility of a single substance dissolved in some solvent which itself may not crystallize along with the dissolved material?

[1] N. L. Bowen, *Jour. Geol.*, Vol. XXV (1917), pp. 218, 235–36, 242–43.

A NEW FORM OF *DIPLOCAULUS*

M. G. MEHL
University of Missouri

To all who have worked with the genus *Diplocaulus*, the great variety in the shape of the skull and the peculiarities in other parts of the skeleton are known. Even after the analysis of the group by Case, Williston, Douthitt, and others the several species assigned to the genus are not entirely satisfactorily defined and there are many details of the anatomy still to be determined.

Recently, through the courtesy of Professor R. D. Salisbury and with the assistance of Mr. Paul C. Miller, the writer was permitted to examine the many specimens of *Diplocaulus* in Walker Museum, the University of Chicago. Among the materials is a recently discovered specimen which, because of its distinctness from described forms and the possible light it throws on the development of the Diplocaulian characteristics, is worthy of description.

Diplocaulus primigenius SP. NOV.

The material herein described consists of a large skull of the type designated by Case[1] as *D. magnicornis*, nine dorsal vertebrae, seven ribs, and a fragment that the writer takes to be a part of the right clavicle. All these parts are well preserved and have been skilfully prepared and mounted on a plaster base by Mr. Miller.

When found the vertebrae formed a curved, but unbroken series extending back from near the posterior border of the skull. At least one of the anterior vertebrae is missing. The ribs lay in an orderly pile to one side of the vertebrae.

THE SKULL

While the skull is essentially complete, its state of preservation prevents a detailed description of its characteristics. Little or

[1] E. C. Case, "Revision of the Amphibia and Pisces of the Permian of North America," *Carnegie Inst. of Washington, Publication No. 146*, (1911), p.21.

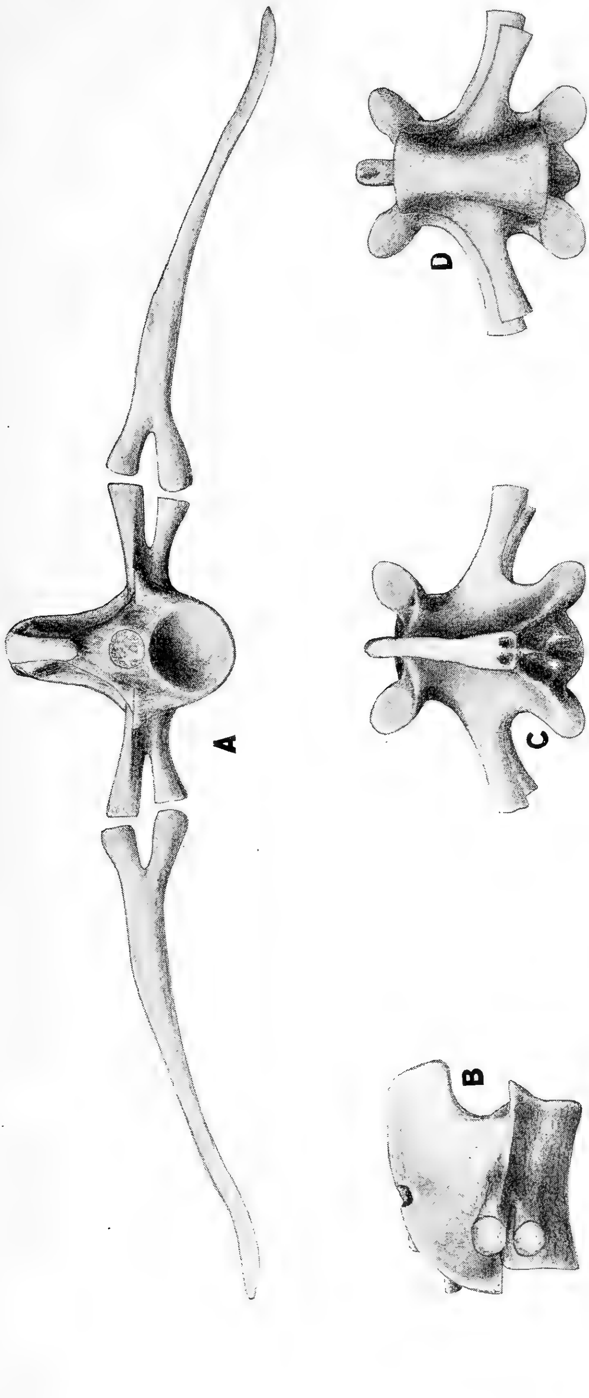

Fig. 1.—*Diplocaulus primigenius*: *A*, anterior view of sixth(?) vertebra with ribs attached; *B*, *C*, and *D*, the same vertebra from the right side, from above, and from below. All figures slightly under $\frac{2}{3}$ natural size.

nothing can be determined concerning those points bearing on the less well-established relations of the various elements. The palate surface has been crushed and lost in part and only the proximal end of the right ramus of the lower jaw is preserved.

A few sutures can be definitely determined, chiefly those of the dorsal surface back of the orbits. In these no departure from the arrangement as shown by Douthitt[1] is noted, but the doubtful portion about the orbits and nares can be neither verified nor disproved. In its general appearance there is nothing to distinguish this skull from any other of a dozen that have been referred to *D. magnicornis*. Although the posterolateral horns are broken off near the tips they are undoubtedly of the bluntly pointed, non-curved variety. The posterior border of the skull is broadly

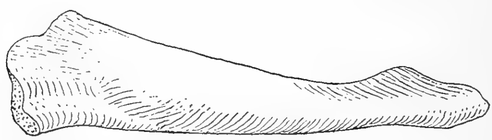

FIG. 2.—*Diplocaulus primigenius:* scapular process of right clavicle from the inner side. Natural size.

concave. None of the details such as the forward extent of the frontal and the arrangement of the vomerine teeth as stressed by Case[2] can be determined. The sculpture of the facial region is decidedly not radial.

The length of the skull along the median line is 116 mm. The tabular horns project back from the posterior border of the skull at the median line a distance of 78 mm. and extend 338 mm. from tip to tip.

The borders of the orbits are largely restored, but the indications are that these openings were above the average in size, possibly as much as 17 mm. or more in diameter.

The possibility of a gill chamber beneath the broad base of the posterolateral horns has been pointed out by Williston, Douthitt, and others. The notch made by the more or less abrupt ending of the

[1] Herman Douthitt, "The Structure and Relationships of *Diplocaulus*," *Contributions from Walker Museum*, Vol. II, No. 1 (1917).

[2] *Op. cit.*, p. 21.

quadratojugal near the mid-length of the horns on their underside apparently forms the anterior border of an opening for the entrance of water into the gill chamber. In most *Diplocaulus* skulls there is a depressed area along the quadratojugal-squamosal union which increases in depth toward the notch, possibly to direct water into the opening. In the specimen herein described the notch is unusually pronounced and the depression leading to it is longer, deeper, and more nearly smooth than is commonly noted.

THE VERTEBRAE

The real interest in the material centers in the vertebrae, especially in their exceptionally large size and in the development of the neural spines. While there is no means of determining accurately the position of the series in the vertebral column, it is assumed that they are from the anterior end, very likely numbers 3 to 11. The first of the preserved series is somewhat shorter than any that follow and it alone has a conspicuous lateral expansion of the neural spine, a condition that suggests its proximity to the skull. Furthermore, this vertebra is the only one of the series that lacks the characteristic pit in the top of the spine, a feature more or less well developed in all the anterior vertebrae except in numbers 1 and 3 in several specimens of the described types examined by the writer.

The string of nine connected vertebrae measures about 354 mm. While this length is made up in a small part of matrix separating the centra, the figure serves well for comparison of this with previously described forms of *Diplocaulus*. The average length of a like number of vertebrae from the same region of the column in several specimens in Walker Museum is but little more than half that given above. All other dimensions of the vertebrae herein described are correspondingly large. So striking is the size that were it not for the fact that in nearly every other detail the vertebrae are those of the typical *Diplocaulus* there might be some doubt as to their identity.

In length the centra increase from about 25 mm. at the anterior end of the series to 40 mm. at the posterior end. The greatest increase is between the first and second, an increase of 5 mm.

The lower side of the centrum is markedly concave anteroposteriorly and distinctly convex from side to side at mid-length.

The diapophysis is somewhat longer than the parapophysis. Both arise from about the mid-length of the vertebrae, the former from the arch and the latter from the centrum. They are united for a short distance at their base. Both increase in diameter toward the distal end where they are distinctly enlarged.

In the described forms of *Diplocaulus* the neural spine has but little development. It is usually little more than a sharp, ridgelike thickening of the arch over the neural canal. In the vertebrae herein described one of the most conspicuous features is the spine development. For the most part the spines are comparatively high with flat, more or less rugose tops. The first of the series is distinctly expanded laterally at its top. It is only in the last two of the series that there is a suggestion of the sharp, keel-like degeneracy of the spine and even in these two vertebrae it rises distinctly above the arch. The ratio between the portion below and above the plane of the zygopophyses throughout most of the column is 4:7 while in the average previously known form the ratio is 4:4. In the last two vertebrae of the present specimen the ratio is about 4:5.

One of the characteristic features of the vertebrae of *Diplocaulus* is the presence of a pitlike depression on the top of the spine. There is a great variety in this pit development ranging from very small, round openings to rather pronounced, laterally elongate depressions. In one string of eleven connected vertebrae, No. 1016 in the Walker Museum Collections, the pits seem to be entirely lacking in all back of the fifth. The first pit is in the second vertebra in every case. Usually it is very conspicuous and more or less quadrangular in shape. The third vertebra apparently lacks the pit. In the following vertebrae the number with pits probably varies from individual to individual. For the most part there is no suggestion of an anteroposterior constriction or division of the pit into two distinct facets and in no specimen except the one herein described has the writer seen distinctly double pitting. In the present specimen, however, in each vertebra except the first of the preserved series which is pitless, the spine depressions are

distinctly paired. In most cases the depressions are connected by a very shallow and narrow groove, but in one or two of the vertebrae the openings are distinct. The size of the pits is another distinctive feature of this form. Between the lateral margins of the pits of the fourth vertebra is a distance of 7 mm. and in every case the pits are so large and so widely separated that the spine is distinctly swollen for their accommodation. Clearly, whatever the function of these pits it was more fully exercised in this than in any of the other forms that have been observed.

As in the typical *Diplocaulus*, the articular faces of the zygopophyses are directed straight up and down so that their common plane furnishes a plane of reference for describing proportions. The zygosphene and zygantrum is strongly developed throughout the series of vertebrae. Posteriorly the neural spine divides to send strong buttresses down, out, and back to the posterior zygopophysis. Between these buttresses there extends back in a horizontal plane a spoutlike projection, the zygantrum. The spine of the preceding vertebra extends forward as a vertical plate running in the groove or spoutlike zygantrum without any noticeable modification for the articulation. There is a suggestion of further strengthening of the intervertebral articulation through a more or less pronounced vertical keel between the posterior, zygopophysial lamellae at the anterior end of the zygantrum excavation. Some of the vertebrae show a corresponding excavation on the lower anterior edge of the zygosphene extension of the spine.

The following table comparing the various dimensions of the present specimen with those of the common type, No. 1018 in the Walker Museum Collections, will serve to emphasize the differences.

	W.M. No. 564 mm.	W.M. No. 1018 mm.
Length of centrum	40	22
Dist. across transverse processes	70	44
Greatest height	40	23
Below plane of zygo	14	12
Above plane of zygo	26	11
Across zygopophyses	36	21

THE RIBS

Of the ribs preserved none resembles the comparatively short, straight, distally expanded form figured by Douthitt[1] in the neck region of his skeletal restoration, and commonly found in the large collections of *Diplocaulus* material. It is entirely possible that none of the typical anterior ribs were preserved. Those present were in a more or less connected pile and not directly articulated with the vertebrae. It is assumed, however, that they belong near the anterior end of the preserved string of vertebrae where they were found, and represent a more primitive stage in the development of *Diplocaulus*. Their exceptional curvature and length are at once striking. Only one is preserved in its entire length. This measures 111 mm. Some of the other ribs were evidently shorter than this but for the most part they indicate an equal or even greater length.

In the one complete rib there is a departure from a straight line of 20 mm., perhaps the average. Some are more sharply curved and others less. The distal half is essentially straight, the curvature being a gradual bending of the proximal half. The curvature is within a plane that departs little from the horizontal so that a body cross-section shows but a slight down-bending of the ribs, a condition entirely in keeping with the generally accepted conception of a broad, flat, bottom-living type. The capitulum and tubercle are both markedly expanded and undoubtedly formed a firm attachment of rib to vertebrae.

HABITS AND RELATIONSHIPS

The aquatic adaptations of *Diplocaulus* have been repeatedly pointed out by various writers. In typical forms there is much evidence that these amphibians were of the groveling, bottom-living type. So bizarre are the skull modifications and body form and so obscure are the stages leading up to this condition that any suggestions are of exceptional interest.

In several respects the form herein described seems to represent an antecedent step in *Diplocaulus* evolution. At least it offers some pointed suggestions as to the several stages through which

[1] *Op. cit.*

the group passed. The large, strong vertebrae; the well-developed spines, and the highly curved ribs approach the normal condition of the more generalized Stegoceph. The well-developed zygosphene and zygantrum articulations bespeak a strength and flexibility of the vertebra column such as belongs to a creature that has developed swift progression through the water by means of tail propulsion.

The remarkable pit development on the spines of this form suggests a still further step in its modifications for swift movement through the water. It seems highly probable that the pits formed an articulation for anteroposteriorly movable spines that gave support to a dorso-median fin. The lateral rigidity of such a fin was assured by the double articulation indicated by the well-developed double pits. As would be expected, the anterior fin spine gives evidence of its superior size and strength of articulation through the very conspicuously larger pit on the second spine. The present specimen lacks the first vertebra, but the several strings of previously known forms examined by the writer establish this point. In the backward folding of the fin spines extra space would be required behind the enlarged anterior spine. Perhaps this accounts for the uniform absence of the pit or pits in the third vertebra.

Not entirely in keeping with these swift-swimming modifications is the remarkable development of the shoulder girdle. The fragment that the writer identifies as the ascending process of the right clavicle is stout and something over 60 mm. long. It is strongly grooved on the outer-posterior side for articulation with the scapula. In all figured specimens of *Diplocaulus* and the many other specimens examined by the writer this process is short. In some cases it may have been broken off but after making due allowances for this it does not seem likely that there was an actual articulation with the scapula or at best, but a weak one. Perhaps the ancestral *Diplocaulus* was a creature that had become well adapted to swift progression through the water to the extent of a suitable vertebral evolution and the reduction of the limbs. At a later stage this actively swimming form likely degenerated to the groveling type in which habitat the girdles continued to reduce

in size and strength of attachment and the vertebral column lost much of its vertical rigidity through the reduction of the spines. Naturally the supports for the dorso-median fin degenerated to a smaller size and a single articular facet. Very likely in the process of flattening the body, one of the first steps was the backward twisting of the curved ribs, a stage preserved in the present specimen, and a later tendency toward straight ribs.

Of the distinctness of this form there can be no question. True, the skull differs very little from many described forms. Greater variations are undoubtedly rightly included in a single species. One might logically expect that the remarkable modifications should reach the skull and limbs before the vertebral column. While all indications are that the skull and associated vertebrae are of the same individual there is, nevertheless, a possibility that the association is of two widely differing species. For this reason the vertebrae and long, curved ribs are considered typical of the new form which the writer wishes to designate as *Diplocaulus primigenius*.

The material on which this form is based is No. 564 of the Walker Museum Vertebrate Collections in the University of Chicago. It was found by Mr. P. C. Miller in Baylor County, Texas, on Brush Creek near Seymour.

The writer takes this opportunity to express his appreciation of the kindness of Professor Salisbury in permitting the study of this material and for the courtesies shown by Mr. P. C. Miller, whose skill and painstaking care in preparing the bones have made it available.

A GLACIAL GRAVEL SEAM IN LIMESTONE AT RIPON, WISCONSIN[1]

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University of Wisconsin

Just west of the city of Ripon (S.E. $\frac{1}{4}$ of N.W. $\frac{1}{4}$, Sec. 20, T. 16 N., R. 14 E.) is a limestone quarry belonging to Mr. William Kroll. This locality (Fig. 1) displays a phenomenon of unusual interest, namely, a seam and pockets of glacial gravel within the Paleozoic rocks.

In 1914 the writer visited this quarry in company with E. O. Ulrich, United States Geological Survey, W. O. Hotchkiss, state

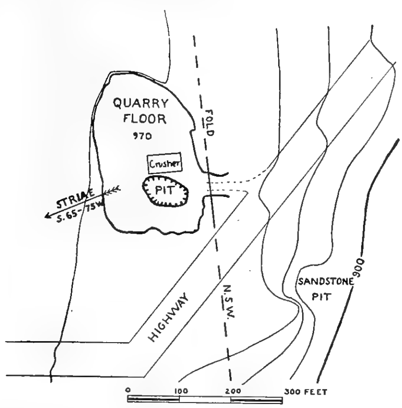

FIG. 1.—Sketch map of Ripon limestone quarry. Topography adapted from U.S. Geological Survey, contour interval 20 feet.

geologist of Wisconsin, and Edward Steidtmann. Further field work was done in 1915 in company with Lawrence Martin and Walter H. Schoewe, and in 1919. The writer is indebted to all of these persons for assistance, criticisms, and suggestions.[2]

[1] Published by permission of the state geologist of Wisconsin.

[2] Lawrence Martin, "The Physical Geography of Wisconsin," *Bull. 36, Wisconsin Geol. and Nat. Hist. Survey,* 1916, pp. 250–51.

A few rods east of the Kroll quarry, on the side of the hill which slopes to the east, is a pit in soft St. Peter sandstone which is overlain, at the top of the exposure, by a few feet of disintegrated Trenton dolomite. Within the sandstone are crevices and pockets filled with glacial gravel. Excavations at the bottom of the pit show the shaly and cherty zone characteristic of the top surface of the underlying Lower Magnesian formation which here dips about 30 degrees to the east.

In the Kroll quarry, on the hill to the west, 50 feet higher than the bottom of the pit, sandstone is seen at the entrance, overlain

FIG. 2.—Fold at entrance of Ripon limestone quarry, looking north

by a few feet of dolomite at about the same level as that to the east. At the eastern edge of the quarry proper is the monoclinal fold shown in Figure 2. The displacement is from 8 to 10 feet, down to the west, and the strike is N. 5° W. Less than 40 feet west of this fold is a pit in the main floor of the quarry 40 feet deep, which shows no St. Peter sandstone. The quarry well reaches sandstone at a depth of 100 feet (Fig. 3).

In Figure 3, No. 1 is Lower Magnesian; Nos. 2 to 4 and 6 to 8, Trenton; and Nos. 5 and 9, Pleistocene; the sandstone in the well is Cambrian.

The gravel seam, No. 5, extends throughout the quarry, which is over 200 feet across. The dolomite layers above the seam are in no way unusual, but along the seam is considerable weathering so that the underlying rock is nearly everywhere disintegrated into a putty-like yellowish clay to a depth of an inch or two. It is this feature rather than the presence of gravel which makes this horizon so prominent, as is shown in Figure 2. The gravel

9. Glacial till	2–3 feet
8. Gray, thin-bedded dolomite	5.0 feet
7. Bluish thin-bedded dolomite	4.5 feet
6. Grayish-blue dolomite	0.8 feet
5. Glacial sand and gravel maximum	0.2 feet
4. Hard gray rather thin-bedded dolomite	5.3 feet
3. Blue, heavier bedded dolomite with fossils; sandy at base	6.5 feet
2. Sandy buff dolomite, forming a parting	0.5 feet
1. Gray dolomite with layers of gray shale and sandstone. Irregular beds, the bedding in general dipping at low angles to the east	40.0 feet
Total, about	63 feet

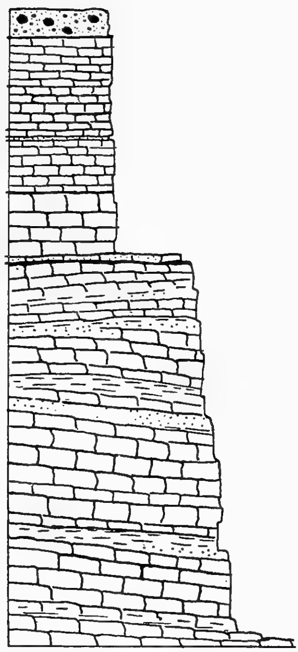

FIG. 3.—Section of Kroll quarry, Ripon, Wis.

and sand layer is best developed in the eastern part of the quarry. In thickness, the sand and gravel layer varies from a mere trace up to a maximum of about two inches (Fig. 4). The largest pebbles are about an inch in diameter. Dolomite, granite, greenstone, and quartzite were distinguished. Minute cross-bedding is locally observed.

In the sandstone pit to the east, larger gravel deposits are found filling gashes or irregular openings in the rock. Some of the openings are as much as three feet across and of irregular shape.

As No. 1 of the section does not contain fossils, the correlation of this section with that seen in the sand pit to the east is somewhat puzzling, but the evidence of the rocks and fossils, of the sloping surface of the Lower Magnesian seen in the sandstone pit, and a close study of the supposed fault unite in definitely proving that the St. Peter pinches out in the interval. Statements of Mr. Kroll regarding exposures formerly visible in a trench through the quarry floor also confirm this explanation. Similar phenomena without the complicating presence of folding are known elsewhere in the vicinity.[1]

FIG. 4.—Close-up view of gravel seam, showing weathering of adjacent dolomite. (Photograph by W. O. Hotchkiss.)

Before considering the origin of the gravel seam, the question of the origin of the fold with which it is related may be considered. The fold might be due:

1. In some way to the pinching out of the St. Peter.

2. To glacial pressure on the east side of the hill, which forced up the wedge of St. Peter sandstone resting upon the sloping surface of the Lower Magnesian.

3. To the same processes which formed the major joints and the other faults and minor folds which are found in the Paleozoic rocks of eastern Wisconsin.

[1] T. C. Chamberlin, *Geology of Wisconsin*, Vol. II (1877), pp. 270–75.

The direction of the movement which caused the fold is not in harmony with slumping of the St. Peter on the inclined surface of the underlying Lower Magnesian. The hypothesis of slumping might, however, explain the eastward-dipping slips and gashes in the St. Peter sandstone.

The forcing up by glacial pressure of a wedge of St. Peter upon the shaly surface of the underlying Lower Magnesian is a satisfactory hypothesis. The strike of the fold is N. 5° W., that of the glacial striae on top of the quarry S. 65–75° W., making an angle of 70 to 80 degrees. Somewhat similar phenomena are described by Chamberlin near Burlington, Wisconsin, and by Alden.[1]

Folds are known in the Paleozoic rocks of the region, as at the abandoned quarry in Fond du Lac, while faults are by no means uncommon. In many instances it can be definitely proved that these features far antedate the glacial period and are not due to ice work. It is therefore necessary to have more evidence than is now at hand to prove that these particular joints and folds at Ripon are different in age from those of the Driftless Area. The localization of the fold by the wedging out of the sandstone will fit with this hypothesis as well as that of glacial pressure.

The writer has therefore concluded that the fold is probably only in part due to glacial action but was principally caused by earth movements.

The following hypotheses may be considered in connection with the origin of the gravel seam between the dolomite layers:

1. Deposition of gravel in Ordovician times between layers of dolomite.

2. Glacial transportation of a slab of dolomite on to a previously planed-off surface, covering a small thickness of gravel.

3. Raising of frozen-together upper layers, possibly during the formation of the fold, followed by filling of the crack with gravel and subsequent settling back of the overlying rocks.

4. Entry of glacial or post-glacial gravel-bearing waters along a bedding plane.

[1] T. C. Chamberlin, *Geology of Wisconsin*, Vol. II (1877), p. 203; Wm. C. Alden, "Quaternary Geology of Southeastern Wisconsin," *U.S. Geol. Survey, Prof. Paper 106* (1918), pp. 206–8.

The absence of similar deposits of proved Ordovician age, together with the lack of induration and the presence of the gravel-filled gashes in the sandstone, definitely eliminate the first hypothesis.

The second theory may be divided into three subdivisions: (*a*) distant source for the dolomite slab; (*b*) movement by rotation without transportation to any great distance; (*c*) slight lateral movement causing the opening of an irregular bedding plane by destroying the registry of the irregularities.

a) No nearby source for such a gigantic slab (at least two hundred feet square) can be found except on the hill northwest of the quarry, and a glacial movement from the northwest is impossible since the Driftless Area is not very far away in that direction. The glacial transportation without breaking of such an enormous piece of rock for any distance, as from across the valley to the east, is quite out of the question. The rock above the gravel seam is no more broken than is usual near the surface of the ground. The major joints, which strike from N. 5° W. to N. 30° W. pass right through the gravel seam (Figs. 5 and 6) and there seems no valid reason to assign a glacial or post-glacial age to them. In fact, there is no difference in the rocks above from those below the seam and no glacial striations are found on the under surface.

b) The second subdivision, namely, the inclusion of gravels by a rotary movement of the overlying rocks, is open to the same objections. It also necessarily involves the assumption of the glacial origin of the fold.

c) The opening of an irregular bedding plane by a slight movement, either rotary or lateral, which destroyed the registry of the irregularities is a possibility. The seam is, however, very regular; moreover, the glacial or post-glacial origin of the joints must be presupposed under this hypothesis.

The third view, the raising of the overlying beds by glacial action thus opening a seam, as one opens the leaves of a book, offers the fewest difficulties. However, it seems impossible that such a process could avoid moving and breaking the overlying rocks. A further difficulty is the comparatively regular thickness of fine gravel. If the frozen upper layers were simply buckled upward

by glacial pressure, as one would buckle a sheet of paper by pushing slightly on one side, the explanation is perfectly adequate. It is not even necessary to assume a glacial origin for the fold, for

Fig. 5.—Gravel seam near entrance to quarry

Fig. 6.—Gravel seam in west face of quarry, showing joints

pressure with the beds in their present position would account for the phenomenon. The filling can be ascribed either to glacial or to post-glacial waters, under this view.

Turning to the fourth hypothesis, it is significant that the gravel seam shows most weathering of all the bedding planes in the quarry, thus indicating that it has been a trunk channel for ground water. It is difficult, however, to understand, first, why the waters have followed only this particular plane; second, how they carried such comparatively large stones considering the size of the opening, and third, how they could produce so widespread and comparatively regular a sheet of gravel without at all affecting

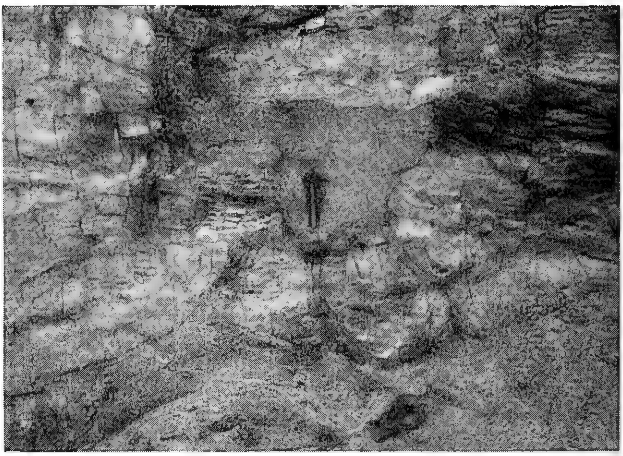

Fig. 7.—Glacial gravel pocket in St. Peter sandstone. (Photograph by W. O. Hotchkiss.)

the large joints which pass through it. Being 10 feet below the surface, the seam is beyond the reach of frost under present climatic conditions unless air circulated through it, so that we cannot appeal with certainty to enlargement by freezing. It is possible that weathering of an originally soft layer has been the cause of lateral enlargement of the bedding plane parting so as to permit a widespread deposit of sand and gravel. In the case of the gashes and joints in the sandstone the effects of erosion by the gravel-carrying waters is clear, while the source of the filling might readily be the gravel deposit which overlies a portion of the exposure. The

position of the gravel seam in the quarry is not such, however, as to lend much support to this.

The writer has concluded that the balance of probabilities favors the view that the gravel seam is due primarily to glacial pressure, although solution has undoubtedly had a part in enlarging the parting. The gravel and sand may well have been in part deposited since the Glacial Period.

STRAND MARKINGS IN THE PENNSYLVANIAN SANDSTONES OF OSAGE COUNTY, OKLAHOMA[1]

SIDNEY POWERS
Tulsa, Okla.

INTRODUCTION

While mapping geologic structure in Osage County, Oklahoma, the attention of the writer was called to certain grooves on bedding planes which resemble glacial striae and slickensides, although clearly of strand origin. They are but one of the many kinds of markings formed along shores or in shallow water. From parallel groovings over broad areas there are all gradations to irregular markings which are clearly casts of impressions in the mud. The markings appear both on upper and lower surfaces of sandstone beds, but they are more common on lower surfaces.

Similar markings in sandy beds of the Portage group of the Upper Devonian of Naples, New York, have been described in detail recently by Dr. John M. Clarke[2] and many years ago by Professor James Hall.[3] They have also been found in the Strawn formation of Pennsylvanian (Pottsville) age, both south of Strawn, Texas, and east of San Saba, in San Saba County, Texas.[4] Grooves of similar nature are known in the Beekmantown of Valcour Island, Lake Champlain.[5] No doubt they are much more common than these few localities indicate.

[1] Published by permission of the director, United States Geological Survey.

[2] "Strand and Undertow Markings of Upper Devonian Time as Indications of the Prevailing Climate," *New York State Mus., Bull. 196* (1918), pp. 199–210.

[3] *Geology of the Fourth District of New York* (1843), pp. 234–37.

[4] The latter locality is known to Dr. J. A. Udden, of Austin, Texas.

[5] Personal communication from Dr. Rudolf Ruedemann, of Albany, New York.

The writer is indebted for information and suggestions regarding these markings to his former associates on the United States Geological Survey, Clarence S. Ross and Kenneth C. Heald. The best localities were found by R. H. Wood and D. D. Conduit of the Survey.

CONDITIONS OF SEDIMENTATION

During the middle and upper Pennsylvanian, when shales, sandstones, and limestones of the Osage were being deposited, a shallow sea with a continually oscillating shore line covered eastern Oklahoma and Kansas, and west-central Texas. The axis of the basin extended in a northeast-southwest direction with the southeastern shore line not far east of the eastern Osage and near the present outcrop of the Strawn formation in Texas.[1] The territory in Oklahoma in which the markings are found lies in the eastern Osage between Pawhuska, Hominy, Skiatook, and Tulsa. In this region limestones are thin and few, but highly fossiliferous. The sandstones commonly show ripple, current, and other strand markings; occasionally they are fossiliferous.[2] Stratigraphically the grooves are known in various horizons between the Hogshooter limestone and the Elgin sandstone, 1,000 feet apart. The Strawn formation in Texas is older than any of the Osage sediments.

DESCRIPTION OF MARKINGS

Nothing unusual has been noted in the current, ripple, and rill marks in the Osage sediments except that the current and rill markings appear in various stages of obscurity and some of the markings defy classification. They appear both on the upper and lower surfaces of sandstone layers. The nature of the parallel groovings may be judged by the photographs.

Localities in the Osage where markings may be seen are numerous. A few are given in the following list: (1) S.E. cor., N.E. $\frac{1}{4}$, Sec. 8, T. 20 N., R. 12 E., near F. A. Gillespie No. 1 (by R. H. Wood and the writer); (2) on

[1] Mr. A. W. McCoy, of Bartlesville, Oklahoma, has prepared extensive manuscript paleogeographic maps of these seas and has kindly given the writer the foregoing information and the correlation of the beds.

[2] "Structure and Oil and Gas Resources of the Osage Reservation, Oklahoma," *U.S. Geol. Survey, Bull. 686* (1919).

a bald hill in W. $\frac{1}{2}$, Sec. 7, same township (by C. S. Ross); (3) N.W. cor., Sec. 6, same township, a V-shaped groove (by C. S. Ross); (4) Sec. 14, T. 24 N., R. 8 E. (by K. C. Heald); (5) N.W. $\frac{1}{4}$, S.W. $\frac{1}{4}$, Sec. 10, T. 25 N., R. 9 E. (by K. C. Heald, D. D. Conduit, and the writer); (6) Sec. 11, T. 21 N., R. 11 E. (by C. S. Ross, P. V. Roundy, and the writer); (7) in the stone wall surrounding the house of Tom Gilchrist and in the foundation of the house of A. P. Kennedy a mile northwest of the Tulsa Country Club, and in a stone house on Duluth Street south of the Country Club (by C. S. Ross and the writer).

The Osage grooves occur on hard, massive sandstones $\frac{1}{2}$ foot to 2 feet in thickness and are seen over broad surfaces or on slabs 1 foot to 5 feet in length. The only pronounced markings, like those illustrated, occur on under surfaces of blocks probably underlain by sandy shale. The grooved block at locality (6) above appears to be in place with rather poor ridges on the upper surface. Near Pawhuska (locality [5]) the markings are faint lines and fainter ridges not over $\frac{1}{4}$ inch in height seen in places on the upper surface of a bed exposed on the top of an unused quarry. They are subparallel and may be found for a distance of 70 feet measured across the ridges, but nowhere for a distance of more than 6 feet along the ridges.

The markings are rounded and smooth. They sometimes have vertical or even undercut sides and characteristically are covered with minute parallel lines on the sides of the larger ridges. Clearly defined V-shaped grooves are infrequent (Fig. 6). The width varies from that of a pencil line to 2 inches; it is uniform until the marking disappears. Neither depth nor height is, as a rule, more than $\frac{1}{2}$ inch, but either may vary quite abruptly or be undulating in any single marking. Single straight markings with a relief of 2 to 3 inches are found alone and also associated with the other strand markings. Their origin apparently is similar to that of the parallel ones. Professor Hall (*op. cit.*) described one cast 6 inches in diameter. The relief of the surface shown in Figure 1, $5\frac{1}{2}$ feet in length across the markings, is not more than 1 inch. No uniformity of spacing between the larger grooves in the cast has been observed. In some cases there are no coarse markings. Small ridges and large grooves in the cast seem to be the rule in one locality, but with the irregular and complicated markings, such as

FIG. 1.—Block of sandstone $5\frac{1}{2}$ feet in height showing parallel strand markings on a bedding plane. This is the lower surface of the bed and the markings are casts of the original. As shown by the ruler in one of the upper grooves, the coarser markings are grooves in this cast. The relief of the surface is less than 1 inch. Locality, S.E. cor., N.E. $\frac{1}{4}$, Sec. 8, T. 20 N., R. 12 E., Osage County, Oklahoma. Photo by Lucian Walker, of Tulsa, Oklahoma.

FIG. 2.—Another block of sandstone at the same locality as Figure 1, showing a curve in parallel markings, and also showing irregular markings which are depressions in the cast. Nothing but flexible objects such as fronds of algae could have produced a curve in some of the markings and not in others when dragged along the strand. The irregular markings near the top may be casts of stranded algae. Photo by R. H. Wood, United States Geological Survey.

Figure 5, the reverse is the case, and there are no depressions in the cast.

Parallelism is the striking characteristic of the markings: they are as straight and as parallel as if ruled with a straight-edge. Yet frequently the grooves end abruptly, but smoothly as seen in Figure 3, or the ridges and grooves in the cast may disappear one by one in a deep, curved, and narrow groove whose beginning is

FIG. 3.—Irregular markings in another block of sandstone at the same locality as Figures 1 and 2. Suspended fronds of seaweed waving back and forth could produce these sinuous, intersecting fine grooves and ridges.

one of the grooves, but whose depth (a sharp ridge in the original) is fully $\frac{1}{2}$ inch. It is a sort of festoon, an arrangement very common in pahoehoe lava where the rumples in the lava are suddenly swept beneath the surface in a smooth curve. Sometimes some of the markings end in a rude curve or are curved along their length, the original direction being resumed beyond the curve (Fig. 2), while the other markings on the same slab are perfectly straight. Associated with some of the grooves are irregular small depressions and less abundant elevations in the cast (Fig. 2,

upper corner, Fig. 4). Cross-markings are infrequent in the well-striated surfaces, but common in the more irregular markings described below (Figs. 3, 6, 7). Professor Hall, however, figured one well-striated slab with three sets of parallel furrows.[1]

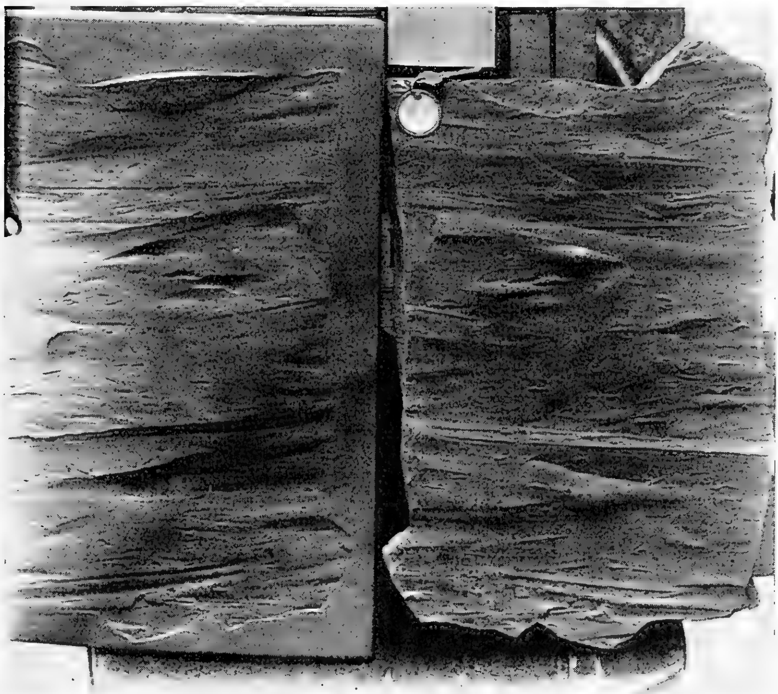

FIG. 4.—A slab of sandstone and a plaster cast of the same showing the original strand surface in the cast. The markings were depressions in the strands. Some of them represent very small dents made by objects swept along, but two of them strongly suggest nodal impressions of plants which, it is believed, lay on the strand. Locality, Naples, New York; Portage group of the Upper Devonian. Specimen in the New York State Museum.

Fossils are occasionally found in the grooves, but small rod-like impressions and gouges in the strand, as seen in Figure 4, are more common. Drifted shells fill grooves in the Beekmantown of Valcour Island, Lake Champlain, according to Dr. Ruedemann. Hall figures a slab showing on the original strand surface, two

[1] Reproduced by Clarke, *op. cit.*, p. 202.

broad ridges separating a broad groove, the latter being filled with drifted shells which washed in after the surface was marked.

Common strand markings have not been noted on the same slab with the most clearly defined grooves and ridges, but well-developed asymmetrical ripple marks, rill markings of various kinds, small irregular markings such as might represent worm

FIG. 5.—Casts of irregular markings, depressions in the strand, which appear to represent fronds of algae which lay on the beach. Near the right lower corner there is a piece of sandstone in the groove which in the original specimen looks suspiciously like the reverse side of a flattened stem of frond. Locality as in Figure 4.

tracks or burrows or algae have been found on sandstones either of the same bed or of beds higher or lower in the series. There are also smooth lumps which might have been lumps of sandy mud, faint single grooves or ridges, and faint symmetrical ripple marks with or without minute, superimposed, tubular lumps and worm tracks, the whole glazed with a thin film of iron oxide as are most of the markings.

Irregular markings, mostly grooves in the casts, are characteristic of many slabs of sandstone associated with the parallel markings. In Figure 3 they have the appearance of being made by objects swept back and forth on the strand while suspended from one or more common centers. Still more irregular markings (Fig. 5), all depressions on the strand, are composed of curved and crossed tubular casts, in cases more than an inch in diameter, and appear to represent casts of algae (or of something similar) which settled in the mud. From this state of irregularity it is but a step to nondescript markings the origin of which is probably inorganic.

Owing to the occurrence of most of the grooves on upturned blocks and to the fact that the beds underlying the casts have not been seen, it is not always possible to determine the original direction of the markings. In Sec. 11, T. 21 N., R. 11 E. it is N. 80° W.; in Sec. 10. T. 25 N., R. 9 E., N. 44°–52° W., which directions are down the depositional dip.

Plant remains have been found in one of the sandstone blocks associated with the grooved blocks in Sec. 8, T. 20 N., R. 12 E. and they occur in sandstone stratigraphically within a few feet of the grooves in Sec. 11, T. 21 N., R. 11 E. Casts of tree trunks have been found in abundance in the section 100 feet from the grooves in the former locality and the very fossiliferous Avant limestone is 80 feet from the grooves in the latter locality.

HYPOTHESES OF ORIGIN

Various hypotheses may be presented for the origin of the grooves: (1) impressions of plant remains, (2) dragging of parts of trees or roots, of algae or of pebbles over the strand, (3) action of ground ice molding unconsolidated sandy clays, (4) differential slipping of the beds before consolidation, (5) tidal action. These hypotheses will be examined in order.

1. *Impressions of plant remains.*—Trunks of trees or branches of algae impressed on the sandy mud could not have made the parallel striations such as are shown in Figure 1. They may well, however, account for some of the single markings, such as Figure 6. Definite worm trails are found on sandstones associated with the

grooved sandstones and casts of branching and interlacing stems generally admitted to represent algae are very common in the Osage rocks. Some of the slabs in the New York State Museum (Figs. 4, 5) show markings which, in the opinion of the writer, were formed as impressions of either algae, or of wood and nodes of plants. In some cases they show faint striations like the plant impressions of the same size on the Joggins section, Nova Scotia, or in the North Sydney section, Cape Breton Island (plates 18 and 22 of Clarke). In other cases the rounded elevations (filling

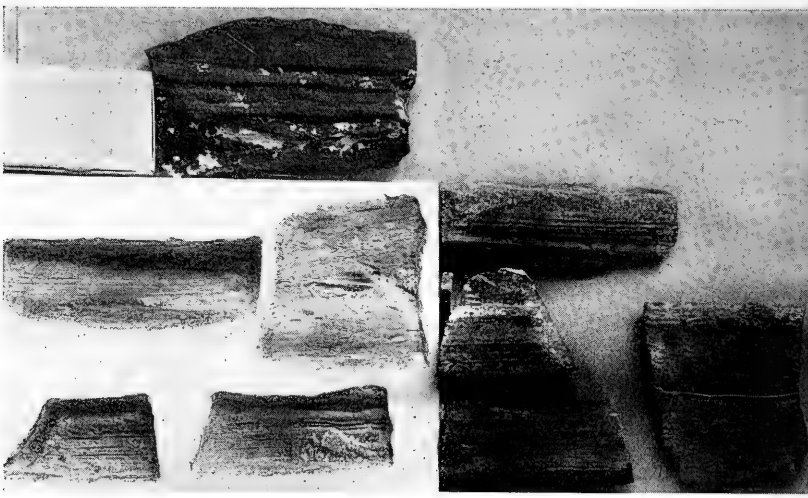

Fig. 6.—Blocks of sandstone in which are casts of V-shaped grooves and which are themselves grooved with sinuous lines. The plaster cast at the left shows the original. A branch or root dragged over fine-grained argillaceous sand could produce such grooves. Locality, Strawn, Texas. The block of sandstone at the top shows somewhat comparable markings produced by weathering of cross-bedding. Locality, basal Eocene, southeast of Uvalde, Texas.

of depressions in the original strand surface) may be partly broken from the rock (Fig. 5), showing that they had an upper as well as a lower surface and may represent casts of kelplike material. In the Tesnus formation of Pennsylvanian (Pottsville) age, 9 miles south of Marathon, Texas, and 225 feet below the top of the formation, there is a heavy sandstone bed showing mud flow and ripple marks and plant casts, but no groovings.

Short, rodlike markings of obscure origin, so-called *Fucoides graphica*, recently have been ascribed to ice crystals by Dr. Clarke at the suggestion of Professor Woodworth.[1] Dr. Udden has enlarged on the suggestion by redescribing similar markings in the Cretaceous of Texas and of the Black Hills as fossil ice crystals.[2] The markings resemble the shorter ones on Figure 4. They are either straight or curved; they are 1 inch to 2 inches in length and $\frac{3}{16}$ to $\frac{1}{4}$ inch in diameter. The same markings are common in the Osage, and in the continuation of the same sandstones near Henryetta, Oklahoma. Faint markings of the same shape occur in the same horizon as some of the grooves. Measured ice crystals have not exactly the same shape as most of the supposed fossil crystals, but the branching and radiating forms are similar. The specimens in the New York State Museum show that the supposed crystals represent casts of crystals or of objects partly buried in the mud, because the ridges in the cast where they join or cross are distinctly superimposed, and in one case the newer object has bent down the older one. Ice crystals could scarcely bend one another, nor could they be depressed by any objects moving over or resting on them. Worm tubes, as

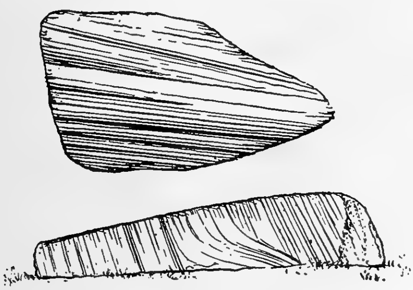

FIG. 7.—Casts of diverging grooves on the lower surfaces of sandstone blocks. The upper figure shows a slab 4 feet long in stone wall near Tulsa Country Club, Tulsa, Oklahoma; the lower figure shows a slab 5 feet long at the same locality as Figure 1.

[1] J. M. Clarke, *op. cit.*, p. 205; J. B. Woodworth, *et al.*, "The Glacial Brick Clays of Rhode Island and Southeastern Massachusetts," *17th Ann. Rept., U.S. Geol. Survey* (1896), Part 1, p. 992; T. M'K. Hughes, "On Some Tracks of Terrestrial and Fresh Water Animals," *Quar. Jour. Geol. Soc.*, London, Vol. XL, No. 157 (1884), p. 184; J. E. Talmage, "Notes Concerning Certain Linear Marks in a Sedimentary Rock" (Abstract), *Quar. Jour. Geol. Soc.*, London, Vol. LII (1896), p. 461; *Univ. Utah Quarterly* (December, 1895).

[2] J. A. Udden, "Fossil Ice Crystals," *Univ. of Texas Bull. (1821)*, (1918); *Sci. Amer.*, Vol. LXXII (1895), p. 102; "A Sketch of the Geology of the Chisos Country, Brewster County, Texas," *Univ. of Texas, Bull. 93* (*Science Series No. 11*) (1907), p. 32; D. D. Christner and O. C. Wheeler, "The Geology of Terrell County, Texas," *Univ. of Texas Bull.* (1918), p. 15.

beautifully exhibited in slabs forming pillars at the south entrance to the Tulsa Country Club (Tulsa, Oklahoma) characteristically are superimposed, and have the same diameter as the supposed crystal tubes. It is barely possible that climatic conditions in the several geologic periods in which these markings were formed were such that ground ice and ice crystals could form.

2. *Dragging of parts of trees or roots of algae or of pebbles over the strand.*—Groovings in sandy clay may be produced by shoving or dragging objects over the surface either above or below water. The alga, *Ulva enteromorpha*, has been observed dragging pebbles over the strand making depressions.[1] Ulva grows in tufts 6 to 8 inches high attached to one side of pebbles as large as $\frac{3}{4}$ inch in diameter. The fronds stand upright from the anchoring pebble, and wave back and forth dragging the pebble, but not lifting it. Either single or double grooves are produced, and the grooves in the illustrations cannot be distinguished from many of the single Osage grooves. One can readily imagine how this operation could produce many of the irregular Osage markings, especially those which cross (Fig. 3), but pebbles are very rare in the Osage sediments.[2] To produce parallel groovings bunches of large algae could be washed along by the tide, and where the algae came to rest irregular depressions would be produced in the sandy shale as stated above. Curves in the grooves, especially in some and not in others, could readily be produced by obstructions in the path of the algae (Fig. 2, upper corner) or by swirls in the tidal action, or by undertow.

Fragments of fossil wood indicate the presence of trees at the same horizon as the parallel markings. Were a log washed over the strand it would normally rotate; were a stump dragged along some of the roots would make deep, sharp-pointed grooves and ridges such as are occasionally found alone (Fig. 6). A log could not make the curved and diverging markings, but light pieces of wood could produce superimposed series of striations. Single grooves with finely striated sides unassociated with other

[1] A. P. Brown, "The Formation of Ripple Marks, Tracks, and Trails," *Proc. Acad. Nat. Sci.*, Philadelphia, Vol. LXIII (1911), pp. 536–47.

[2] C. S. Ross reports a conglomeratic limestone in the south part of T. 22 N., R. 7 E., that contains gravels and small pebbles of white vein quartz, igneous, and metamorphic rocks which were possibly carried by floating ice.

strand markings have the appearance of being made by roots or other portions of trees.

3. *Action of ground ice molding unconsolidated sandy clays.*—Floating land ice has been shown by Lyell to be capable of furrowing consolidated sandstone.[1] Dr. Clarke and Professor Woodworth concur in the opinion that the markings in the Upper Devonian of New York are of similar origin. The occurrence described by Lyell was at Cape Blomidon, on the Bay of Fundy, where heavily "packed" ice often 15 feet thick with fragments of amygdaloidal basalt frozen in the base, is pushed over ledges of Triassic sandstone with the rise of the tide.

Climatic conditions such that land and ground ice could be formed at various times during the deposition of 1,000 feet or more of sediments in the middle and upper Pennsylvanian are within the range of possibility and the faunas and floras would not necessarily prove or disprove such an assumption.[2] But if floating ice were present it should have gouged out the muds in some places and produced considerable lumps in others. Contrast the absence of such disturbances with Dr. Kindle's description of the action of "ice-shoved bowlder or ice cake" in the Mackenzie River:

> The plowing and gouging action of ice is nearly everywhere in evidence along the Mackenzie. At the head of the river, in the eastern shallow channel, one can see through the clear water numerous deep grooves made by ice cakes or bowlders pushed by ice in the bowlder clay of the bottom. In the gravel or slits of low islands the broad grooves made by ice-shoved bowlders or ice blocks can often be traced for a considerable distance. In some localities the plowing and scooping action of the ice carries quantities of mud from the bottom to the banks of the river.[3]

A comparatively smooth surface on the bottom of the ice would produce the required parallel markings and might produce cross-markings, but not the irregular markings unless small cakes of

[1] *Travels in North America*, Vol. II (1845), p. 144.

[2] "It seems possible to state that there is evidence for presuming that the Permian glacial period was preceded in the Carboniferous by a degree of cold permitting floating ice in continental bodies of water and also in the sea in middle latitudes."—J. B. Woodworth, "Boulder Beds of the Caney Shales at Talahina, Oklahoma," *Bull. Geol. Soc. Amer.*, Vol. XXIII (1912), pp. 457–62; quotation p. 462.

[3] "Notes on Sedimentation in the Mackenzie River Basin," *Jour. Geol.*, Vol. XXVI (1918), pp. 341–60; quotation p. 353.

ice floated or were dragged around. Some of the irregularities (largely sharp ridges and elevations in the original) might have been formed by the melting of ice around mud-filled cracks. There was no regional Pennsylvanian glaciation in the Osage, for tillites have not been found. Bowlders of large size which have been interpreted as indicating the presence of floating ice during the early Pennsylvanian (Pottsville) in the vicinity of the Arbuckle and Ouachita mountains has caused considerable discussion.[1] Recently Professor Samuel Weidman has found striated pebbles in the Franks conglomerate of the Arbuckle Mountains together with the striated floor of older rocks on which the conglomerate rests.[2] This conglomerate is much older than the sediments of the Osage. Another reported locality of tillite near Shafter, Presidio County, Texas, has been examined by the writer and is believed by him not to be tillite.[3]

Opposed to the theory of floating ice is the fact that nowhere are any of the striated sandstone beds gouged or beveled obliquely to the bedding planes; nor are the latter in any way disturbed. The varied and complex strand markings, which are much more abundant than the grooves on the various blocks of sandstone at the same locality, have never been rubbed by ice.

4. *Differential slipping of beds before consolidation.*—Slipping in unconsolidated beds, especially in muds and silts, is a well-known phenomenon recently emphasized in the mud lumps of the Mississippi delta[4] and in experiments by Dr. Kindle.[5] The latter

[1] J. A. Taff, "Ice-borne Boulder Deposits in Mid-Carboniferous Marine Shales," *Bull. Geol. Soc. Amer.*, Vol. XX (1909), pp. 701–2; *Science* (New Series, Vol. XXI (1905), p. 225; E. O. Ulrich, "Revision of the Paleozoic Systems," *Bull. Geol. Soc. Amer.*, Vol. XXII (1911), p. 352, footnote; J. B. Woodworth, *op. cit.*

[2] "The Probability of Pennsylvanian Glaciation of the Arbuckle Mountain Region," *Bull. Geol. Soc. Amer.*, Vol. XXXII (1921).

[3] J. A. Udden, "The Geology of the Shafter Silver Mine District, Presidio County, Texas," *Univ. of Texas Mineral Survey, Bull. 8* (1904), pp. 14, 59. The only conglomerates found by the writer are at the base of the Cieneguita series as determined by the intrusive contact with syenite porphyry. They are very fine conglomerates and sandstones composed of rounded quartz pebbles and grains which look like fragments of vein quartz rounded by long water erosion.

[4] E. W. Shaw, "The Mud Lumps at the Mouths of the Mississippi," *U.S. Geol. Survey, Prof. Paper 85b* (1914), pp. 11–27.

[5] E. M. Kindle, "Deformation of Unconsolidated Beds in Nova Scotia and Southern Ontario," *Bull. Geol. Soc. Amer.*, Vol. XXVIII (1917), pp. 323–34.

has shown that when soft beds are overlain by unequally disturbed heavier and firmer beds there is a marked deformation of the former caused by differential weighing of the firmer upper beds. In the Osage the overlying heavy sandstones are equally distributed over large areas, but were the strands inclined, slipping could conceivably take place in sediments of the right composition producing modified slickensides on the bedding planes.[1] Again, were there readjustments within the subsiding geosynclinal basin during deposition, such as faulting, which Mr. A. W. McCoy tells the writer took place in the Osage basin, slipping would be expected along bedding planes, just as slipping is known to have been caused by earthquakes.

Difficulties not overcome by a slipping hypothesis, even assuming that sandy clays and sandstones would slide, are the explanations of systems of cross-markings, and the absence of buckling of the strata or of intraformational conglomerate. Slickensiding after consolidation is not a possibility because the markings are clearly molded, not gouged.

5. *Tidal action.*—Current and tidal action produce a variety of markings on the strand, a number of which have been figured by Dr. Clarke and which can be reproduced in the Osage sandstone and in the Tesnus formation sandstone of the Marathon region, Texas; especially the "mud flows" which are interpreted as plunge and undertow markings of retreating waves on the beach. The larger ones, which are 2 to 5 inches in width and 1 inch in depth may have been formed, as suggested to the writer by Mr. C. A. Hartnagel, of the New York State Museum, in tiny puddles of water on a gently sloping beach swirled by sudden gusts of wind. Professor Hall's theory of the origin of the parallel markings is by current action:

> The only assignable cause for these ridges is the action of a current flowing over the surface of the strata, sometimes transporting sand and at other times coarser material which furrowed the surface upon which the subsequent deposits were made.[2]

[1] Dr. J. A. Udden has kindly called the attention of the writer to photographs of sandstone slabs showing minute faults and wavy subparallel folds, the origin of which he believes to have been creep of settling muds before consolidation, but these folds do not resemble the strand markings (*Univ. of Texas, Bull. 246* (1912, Pl. 25).

[2] *Op. cit.*, pp. 234–35.

Cross-bedded sandstones may show very similar grooves and ridges when relief is accentuated by weathering (Fig. 6), but the two kinds of markings are entirely distinct in origin.

CONCLUSIONS

Parallel groovings on bedding planes such as described above have been shown to be markings on the original strand either in sandy clay or in a fine-grained sandstone. Preservation of either the original surface, as in the case of the sandstones, or of a cast of the surface, as in the case of the sandy clays, has failed to reveal both original and cast of any one set of markings except in the form of small fragments.

Inspection of the various theories which may be advanced to account for the formation of such strand markings shows that they have been formed by the dragging of plants, probably stems and fronds of algae, over sand and mud in shallow water, probably by tidal or undertow currents. Ground ice and shore ice are shown to be incapable of having made these markings.

Reasons for the conclusions above may be briefly summarized: (1) general distribution through formations of similar origin, but of different ages in various regions; (2) perfect parallelism over broad surfaces with notable undulations and sinuosities confined to certain of the markings; (3) curved, overlapping, and cross-striations; (4) absence of any disturbance within the strata; (5) intimate association with beach and shallow water markings as current, ripple, rill marks, worm borings, and algal impressions; (6) occasional association with heaps of shells found in the grooves and stratigraphic proximity of plant remains; (7) negative evidence of the presence of ice in any form or of glacial deposits in the containing formations; (8) complete gradation from straightness to curvature to cross-markings and more irregular markings showing nodal and stemlike impressions; and (9) the outstanding delicacy of the whole, showing that gouging was subordinate to molding by objects of light weight.

SUMMARIES OF PRE-CAMBRIAN LITERATURE OF NORTH AMERICA

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V. THE EASTERN PART OF THE UNITED STATES

During the period covered by these summaries, the following United States Survey quadrangle areas containing pre-Cambrian rocks have been mapped: the Raritan quadrangle of New Jersey, the Tolchester quadrangle of Maryland, and the Ellijay quadrangle of Georgia. The New York State Museum has published several papers on Adirondack areas. The Federal Survey has published Emerson's bulletin on the "Geology of Massachusetts and Rhode Island."

One of the most notable advances is the determination of the pre-Cambrian age of the Wissahickon mica gneiss of southeastern Pennsylvania by Bliss and Jonas. In Vermont, the unconformable contact between Cambrian and pre-Cambrian has been more clearly defined by Dale and by Keith. The Ocoee group in the south is now placed by the Federal Survey with the Cambrian. This is still largely a matter of arbitrary decision. The apparently conformable gradation of Ocoee into Cambrian is interpreted by Keith and associates as evidence of the Cambrian age of the Ocoee. The lack of fossils in the Ocoee has been emphasized by Van Hise and Leith as a pre-Cambrian trait.

The Raritan quadrangle[1] lies in both the Appalachian and Coastal plain provinces of northern New Jersey. For purposes of mapping, the pre-Cambrian rocks of the quadrangle are classified as the Franklin limestone, Pochuck gneiss, and graphite schists, all of sedimentary origin. The Byram gneiss is a gray granitoid igneous rock composed of microcline, microperthite, quartz, and

[1] W. S. Bayley, R. D. Salisbury, and H. B. Kummel, "Description of the Rariton Quadrangle, New Jersey," *U.S. Geol. Surv., Geol. Atlas,* U.S. Raritan Folio (No. 191) (1914). 32 pp., 21 figs., 5 maps, section sheet.

hornblende with a little pyroxene and biotite. The constituents of the Losee gneiss are quartz, oligoclase, pyroxene, and some hornblende and biotite. The Pochuck gneisses are mostly of unknown origin, but in part are igneous. They are dark-colored rocks composed of pyroxene, hornblende, oligoclase, and magnetite. The stratigraphic relations of the pre-Cambrian rocks have not been worked out. Their classification is lithological. They are believed to be related to the Grenville series of the Adirondacks and southeastern Canada.

Bayley[1] reports that the pre-Cambrian rocks of the highlands of New Jersey include a series of limestone, quartzites, conglomerates, slates, and micaceous schists whose stratigraphic succession is uncertain. They are surrounded by older and in part igneous gneisses.

Bliss and Jonas[2] conclude that the Wissachickon mica gneiss of the Doe Run and Avondale region of southeastern Pennsylvania is of pre-Cambrian age and is separated by a thrust fault from Ordovician limestone.

Cushing and Ruedemann[3] describe the Saratoga Springs area which lies in the eastern central portion of New York state. It includes portions of the Adirondack highlands, New England plateau and the Champlain downwarp. The pre-Cambrian rocks include Grenville sediments intruded by Laurentian granite. Later intrusions of anorthosite, syenite, granite, and gabbro followed in the order named. The Grenville sediments consist chiefly of a variety of schists probably representing metamorphosed muds. Associated with them is a belt of quartzite with some limestone lenses. The schistosity and bedding of the sediments are inferred to be parallel. The schistosity strikes east and west and dips southward at a low angle rarely reaching 40°.

[1] W. S. Bayley, "The Pre-Cambrian Sedimentary Rocks in the Highlands of New Jersey," *Congrès Geologique International.* (XII Session Canada, 1914, pp. 325-34.)

[2] Eleanora F. Bliss and Anna I. Jonas, "Relation of the Wissahickon Mica Gneiss to the Shenandoah Limestone and Octoraro Schist of the Doe Run and Avondale Region, Chester County, Pennsylvania," *U.S. Geol. Surv., Prof. Paper 98* (1916), pp. 9–34, 3 pls., 3 figs.

[3] H. P. Cushing and H. Ruedemann, "Geology of Saratoga Springs and Vicinity," *New York State Museum, Bull. No. 169*, 177 pp., 17 figs., 3 maps.

Dale[1] traced the boundary of the pre-Cambrian and the Cambrian rocks of Vermont for a distance of 60 miles and finds them to be structurally discordant and unconformable. The pre-Cambrian rocks include various granite gneisses, aplite gneiss, metamorphic arkoses, quartzite, conglomerate with pebbles of quartzite, albitic sericitic schists, and graphitic sericitic schist.

Eaton[2] states that the pre-Cambrian rocks of South Mountain, Pennsylvania, near 40° 20′ north latitude and meridian 76° 10′ west longitude, consist mainly of granite, diorite and gabbro gneisses cut by granite pegmatites. These gneisses probably correspond in age and composition to the Losee, Byram, and Pochuck gneisses of eastern Pennsylvania and New Jersey.

Emerson[3] recognizes two belts of pre-Cambrian rocks in Massachusetts, a western belt forming the backbone of the Green Mountains, the eastern belt extending from Rhode Island through Worcester and Essex counties, Massachusetts. The oldest rock in the western belt is the Hinsdale gneiss, a coarse granitoid gneiss including beds of limestone, quartzite, micaceous graphitic schists. Coarse feldspathic rocks locally replace the limestones. Hornblendic and fibrolitic rocks are also included in the Hinsdale gneiss. In the upper portion of the Hinsdale gneiss is the Cole Brook limestone, a coarse magnesian limestone, highly metamorphosed and about 600 feet thick. A more quartzose gneiss than the Hinsdale is called the Washington gneiss. The dominantly igneous pre-Cambrian rocks of western Massachusetts include the Stanford granite gneiss, titanite-diopside, diorite aplite, Lee quartz diorite, Becket granite gneiss, and dunite.

The oldest pre-Cambrian rocks in the eastern belt is the Northbridge granite gneiss. With apparent unconformity, it is overlain successively by the Westboro quartzite and the Marlboro formation, both doubtfully pre-Cambrian. The latter is a biotite schist.

[1] T. Nelson Dale, "The Algonkian-Cambrian Boundary East of the Green Mountain Axis in Vermont," *Am. Jour. Sci.*, 4th Ser., Vol. XLII (1916), pp. 120–24, 1 fig.

[2] H. N. Eaton, "The Geology of South Mountain at the Junction of Berks, Lebanon, and Lancaster Counties, Pennsylvania," *Jour. Geol.*, Vol. XX (May–June, 1912), pp. 331–43, 2 figs.

[3] B. K. Emerson, "Geology of Massachusetts and Rhode Island," *U.S. Geol. Surv., Bull. 597* (1917), 289 pp., 10 pls., 2 figs.

Fenner[1] advocates the theory of the origin of certain gneisses by injection.

Katz[2] tentatively assigns certain quartzites, slates, and schists of southwestern Maine to the Algonkian because of their lithologic resemblance and area and structural relationship to the Westboro quartzite and Marlboro formation of eastern Massachusetts.

Keith[3] has traced an unconformity at the base of the Cambrian along the west border of the Green Mountains, and concludes that certain older sediments beneath the unconformity are properly classed as Algonkian.

La Farge and Phalen[4] follow Keith in placing the Ocoee group of the southern Appalachians in the Cambrian. In the Ellijay quadrangle of northern Georgia they recognized several groups of pre-Cambrian rocks, all of which they classify as Archean.

The most abundant types comprise an older complex of acid schists and gneisses whose origin is doubtful, and a younger group of areally less extensive basic gneisses and schists, mostly dioritic gneiss which is intrusive into the older complex. The first is known as the Carolina gneiss, the latter as the Roan gneiss. Intimately associated with the Roan gneess, are small masses of pyroxenite and dunite which are probably intruded into the Roan gneiss. Both the Roan and the Carolina gneiss are intruded by small masses of granite believed to be Archean in age.

Martin[5] recognizes a Grenville series and post-Grenville intrusives in the Canton quadrangle of northern New York. The Grenville includes limestones, garnet, and siliceous gneisses, quartzites and quartz schist and amphibolite. The post-Grenville

[1] C. N. Fenner, "Mode of Formation of Certain Gneisses in the Highlands of New Jersey" (Abstract), *Geol. Soc. Am. Bull.*, Vol. XXV, No. 1 (March 30, 1914), pp. 44–45.

[2] F. J. Katz, "Stratigraphy in Southwestern Maine and Southeastern New Hampshire," *U.S. Geol. Surv., Prof. Paper 108* (1918), pp. 165–77.

[3] A. Keith, "A Pre-Cambrian Unconformity in Vermont," *Geol. Soc. Am. Bull.*, Vol. XXV, No. 1 (1914), pp. 39–40.

[4] L. La Farge and W. C. Phalen, "Georgia, North Carolina, Tennessee," *Ellijay Folio, No. 187* (1913), 17 pp., 4 maps.

[5] James C. Martin, "The Pre-Cambrian Rocks of the Canton Quadrangle," *New York State Mus., Bull. No. 185* (1916), 112 pp., 20 pls., 31 figs., maps.

intrusives listed are gabbro-amphibolite, granite gneiss, and pegmatite dikes.

Miller[1] gives the following classification of the pre-Cambrian rocks in the region of Bethlehem, Pennsylvania:

Algonkian	Franklin limestone Vera Cruz graphitic schist
Undifferentiated pre-Cambrian	Acid and basic igneous and sedimentary gneisses cut by dikes of basalt and pegmatite

Miller[2] and others describe the pre-Cambrian rocks of the Tolchester quadrangle east of Baltimore. The pre-Cambrian rocks include the acid Baltimore gneiss and the Wissahickon gneiss, both believed to be largely sedimentary. Their age relations are uncertain. These rocks are intruded by pre-Cambrian granite, gabbro, peridotite, and pyroxenite.

Miller[3] ascribes the foliation of the Grenville series of New York mainly to recrystallization caused by heat and pressure accompanying the upwelling of magmas, and only to a very minor degree to lateral compression. Low dips, parallelism between bedding and foliation, and general absence of small folds are the principal facts on which this view is based. The foliation of the granite syenite series, he thinks, is an original flow and crystallization structure. The same view is taken of the granulated anorthosite and gabbro phases.

Peck[4] states that the pre-Cambrian rocks of Chestnut and Marble hills in Northampton County, Pennsylvania, consist of a lower granitoid, gneissose series, overlain by a highly metamor-

[1] B. L. Miller, "The Mineral Pigments of Pennsylvania," *Pennsylvania Topog. and Geol. Surv., Rept. No. 4* (1911), 101 pp., 29 pls., 9 figs.

[2] B. L. Miller, E. B. Mathews, A. B. Bibbins, and H. P. Little, "Description of the Tolchester Quadrangle, Maryland," *U.S. Geol. Surv., Geol. Atlas,* Tolchester Folio (No. 204) (1917), 15 pp., 3 pls., maps and illus., 3 figs.

[3] W. J. Miller, "Origin of Foliation in the Pre-Cambrian Rocks of Northern New York," *Jour. Geol.,* Vol. XXIV (1916), pp. 587–619, 1 fig.

[4] F. B. Peck, "Preliminary Report on the Talc and Serpentine of Northampton County and the Portland Cement Materials of the Lehigh District, Pennsylvania," *Pennsylvania Topog. and Geol. Surv., Rept. No. 5* (1911), 65 pp., 17 pls. (incl. geol. map), 9 figs.

phosed series of rocks which vary widely in character, and include beds of limestone and dolomite.

Wherry[1] states that the pre-Cambrian rocks of Pennsylvania occur in three distinct belts: (1) the Catoctin belt extending southwest from Harrisburg into Maryland; (2) the Highland belt extending from a point about 40 miles east of Harrisburg and crossing the Delaware River at Easton; (3) the Piedmont Belt which stretches from Philadelphia to Trenton, New Jersey. About half of the pre-Cambrian rocks of the Highland belt are of sedimentary origin. The latter include crystalline limestones, quartz, mica schists, graphite-bearing quartzite, and amphibolitic gneiss, the latter being areally the most important. The principal facts on which belief in sedimentary origin of these rocks is based, include high silica and alumina content, high carbonate content, rounded zircons, the great longitudinal extent of the gneiss laminae, and the greater age of the laminae as compared with granitic intrusions of the region.

[1] E. T. Wherry, "Pre-Cambrian Sedimentary Rocks in the Highland of Eastern Pennsylvania," *Geol. Soc. Am. Bull.*, Vol. XXIX (1918), pp. 375-92.

[*To be concluded*]

EDITORIAL NOTE

In Number 6, Volume XXVIII (1920), of the *Journal of Geology*, it was announced that, because of the extremely high cost of printing, it would be necessary to limit Volume XXIX (1921) to six numbers. This announcement was made with great regret. It is therefore with correspondingly great pleasure that the *Journal* now is able to make the announcement that, through the generosity of one of its associate editors, Dr. R. A. F. Penrose, Jr., formerly connected with the Department of Geology in the University of Chicago, the announced reduction in the volume will not be necessary. During the year 1921, the *Journal* will continue to be published semi-quarterly, as heretofore. The editors of the *Journal* and the University appreciate deeply the generous support of Dr. Penrose. Their feeling will be shared, we believe, by all who are interested in the science of geology.

Certain changes in the editorial policy of the *Journal* have been adopted. These are printed near the bottom of the inside page of the first cover of this number, and the attention of contributors is called to them.

Reviews

Fifteenth Biennial Report, Colorado Bureau of Mines, for 1917 and 1918. Denver: The State Printers, 1919.

The mining is considered by counties and by products. The history, recent development, production, and markets for the various ores are discussed. Non-metallic products are included; also a short note on oil shale possibilities.

In general the report shows that the mining industry of the state is declining. Since 1915 the production of gold in the Cripple Creek district, the chief gold center of the state since 1893, has decreased from $13,683,494 to $8,300,000 (estimated) in 1918. The production of silver in Lake County (the leading silver-producing county) has fallen from 4,154,913 ounces in 1907 to 2,353,530 ounces (estimated) in 1918, although 1914 was a relatively good year. The production of lead has decreased less than that of the precious metals, but the decrease in both copper and zinc has been considerable in recent years. Lake County produces more silver, lead, copper, and zinc than any other. In 1916 the state produced nearly $5,000,000 worth of tungsten, but the estimate for 1918 is less than half this figure, due to decreased demand and possibly to the irregularity of the veins.

Colorado leads the world in the production of molybdenum, the main deposit (said to be the largest known) being in the western part of Summit County. In 1918 the state had an estimated production of 94,000 pounds of uranium, the largest except in 1914. Two million pounds of vanadium (largest production to date) is the estimate for 1918.

The total mineral production of the state to 1917 is as follows:

Gold	$623,047,160	....
Silver	466,463,217	593,796,442 fine ounces
Lead	173,909,020	3,962,140,896 pounds
Copper	35,755,138	237,422,282 pounds
Zinc	106,310,030	1,484,929,849 pounds
	$1,405,484,565	

D. J. F.

Coals and Structure of Magoffin County, Kentucky. By ILEY B. BROWNING and PHILIP G. RUSSELL. Frankfort: Kentucky Geological Survey, 4th Series, Vol. V, Pt. II, with geologic ection and maps, 1919. Pp. x+552.

This is a detailed report on the subject named in the title. The columnar section accompanying the report shows twenty-three beds of coal, not all workable, most of which are in the Pottsville Series. It is stated that only three horizons in the 1,200 foot section are sufficiently persistent and well defined to be serviceable as horizon markers. It is stated that all the strata exposed are of marine origin.

R. D. S.

Oil and Gas Resources of Kansas. By RAYMOND C. MOORE and WINTHROP P. HAYNES. Lawrence: State Geological Survey of Kansas, Bulletin 3. 391 pages, 40 plates.

The volume contains a historical sketch of the oil and gas industry of the state, and brief discussions of a general nature on (1) the origin of oil and gas, (2) their migration and accumulation, and (3) methods of production, refining, etc. These discussions are followed by a summary of the stratigraphy of Kansas (pp. 78–173), including the fullest account to date of the sub-surface crystalline rocks of the state. These rocks (granite) are said to constitute a buried ridge nearly 175 miles long and 10 to 25 miles wide, trending in a northeast-southwest direction (really north-northeast, south-southwest) from the Nebraska line near Bern, to northern Butler County. Its highest elevation is at the north, where its top is about 600 feet below the surface, and its maximum height above the surrounding crystalline rock floor probably is 2,500 feet or more. The age of the granite is conjectured to be pre-Cambrian, and to have been uplifted in the late Mississippian or early in the Pennsylvanian.

These preparatory chapters precede the main topic of the bulletin, the production of oil in Kansas (pp. 194–397). Most of the oil of the state is from the Pennsylvanian system, but the Permian, and perhaps the Mississippian, have yielded some. The production of oil in 1916, the last year for which data are given, was about 8,750,000 barrels, more than twice that of any preceding year. In 1916 more than 3,600 new wells were completed, about 10 per cent of them dry.

A small but clear geological map of the state accompanies the volume, also a map showing the distribution of oil and gas.

The volume bears no date on title-page, or elsewhere where a date is naturally looked for, though the date 1917 appears under the state printer's name. Its publication appears to have been delayed, as so many other volumes have been in recent years.

R. D. S.

Petroleum and Natural Gas in Indiana. By W. M. LOGAN, State Geologist. Fort Wayne: The Department of Conservation, Division of Geology, 1920. Pp. 279.

Like the preceding, this volume appropriately discusses the general fundamental questions concerning the origin and accumulation of oil and gas, and methods of finding it (pp. 10–48). A summary of the stratigraphy of the state (pp. 50–62) is followed by reports on the several counties. A map showing the oil and gas areas of the state accompanies the report.

R. D. S.

The Sand and Gravel Resources of Missouri. By C. L. DAKE. Rolla: Missouri Bureau of Geology and Mines. Vol. XV, 2d ser. (1918). Pp. xii+274, 17 plates.

A useful volume, dealing not only with the geological phases of the subject, but with the industrial phases as well. It is not restricted to surface sands and gravels, but includes available materials of these types in formations from the Cambrian up. Incidentally the volume presents a brief, up-to-date summary of the stratigraphic succession of the state, which is welcome and useful. The volume should be of value to those engaged in most sorts of construction work, both now and in the future, as well as to geologists.

R. D. S.

The Physical Features of Anne Arundel County. By HOMER P. LITTLE and OTHERS. Baltimore: Maryland Geological Survey, 1917. Pp. 232, pls. 9.

This county report covers the physiography, geology, mineral resources, soils, climate, magnetism, and forests. The county lies in the coastal plain, and formations older than the Cretaceous therefore are wanting. One of the striking features of the geology of the region is the large number of unconformities in the Coastal Plain series. There

are, for example, seven Cretaceous formations, each bounded above and below by an unconformity. Much the same may be said of the later formations. The Cretaceous strata of the region have a total thickness of 720 feet, the Eocene, 160 feet, the Miocene, 100 feet, the Pliocene (?), 40 feet, and the Pleistocene about 100 feet.

R. D. S.

Onaping Map-Area. By W. H. COLLINS. Ottawa: Canadian Geological Survey, Memoir 95, 1917. Pp. viii+157, pls. 11, figs. 8, map.

A very concise report on the geology of an area of approximately 3,500 square miles the center of which is 50 miles north of Sudbury. The area lies within the southern part of the pre-cambrian shield and its topography is that of a hummocky plateau 875 to 1,450 feet above the sea. The most important physiographic features antedate glaciation. The two intersecting series of parallel lake basins, in the southwest quarter of the area, probably follow faults.

The solid rocks, all pre-Cambrian, are separable by a great unconformity into a pre-Huronian group and a Huronian group. The pre-Huronian consists of a schist-complex and intrusive granite-gneisses. The schist-complex consists of volcanics and subordinately of water-deposited tuffs, iron-formation, and other sediments. The structure of this schist-complex, wherever determinable, is that of low anticlinoria and synclinoria. Dynamic metamorphism has converted the original volcanics and sediments into chlorite and sericite or paragonite schists. Near the granite-gneiss batholiths the effects of contact metamorphism are very marked. This schist-complex represents a period of extensive vulcanism and the formation of shallow-water or land deposits. The intrusive granite-gneiss series is dominantly granodiorites with which are associated a great variety of amphibolites, diorites, aplites, pegmatites, and other types. The diversity of types is explained by primary differences in the intruding magma, magmatic differentiation, and large-scale magmatic assimilation of older rocks. Crenulated interlocking contacts of larger mineral individuals with irregular shape and orientation are textural features very characteristic of these assimilated products. Good photomicrographs are shown to illustrate these features. In the future these criteria may prove of great assistance in determining this obscure type of metamorphic rocks.

The Huronian rocks constitute the Cobalt series, divisible into two parts. The lower part (Gowganda formation, 0–3,000 feet thick)

is composed of conglomerates, greywackes characterized by incomplete weathering and imperfect sorting, and a few beds of limestone. By many geologists familiar with this general region this formation in part at least is thought to be of Glacial origin. The upper part of the series consists of quartzites (chiefly Lorrain quartzite). As compared with the pre-Huronian, the Cobalt series is little metamorphosed or folded. In most places the Gowganda formation grades up into the Lorrain quartzite, but at some localities there is evidence of an erosion unconformity between the two. This local unconformity may be the result of overlap and probably does not represent a great time-gap.

Both the pre-Huronian and Huronian are intruded by dikes and sills, probably of Keweenawan age. Many different rock-types ranging from norites to aplites are represented and here again the field evidence and relationships make it clear that this diverse petrological variety is due in some cases to original differences in the composition of the magma, in others to assimilation of country rocks, or to magmatic differentiation. Calcite and the association of quartz, chalcopyrite, and silver-cobalt-nickel minerals, which constitute the silver-cobalt veins of the area, are believed to be among the subsidiary differentiates. Primary calcite is found sparingly in the diabase dikes and abundantly in the aplite dikes. In two cases the aplite dikes merge into calcite veins. These sills and dikes and all older rocks of the region are cut by porphyritic olivine diabase dikes.

The numerous gold-quartz veins near West Shiningtree Lake are irregularly mineralized and the gold content is low. In this general region the post-Cobalt diabases have gold-bearing quartz veins associated with them. The gravels along the Vermillion river have been worked for placer gold, but they are rather lean. Small silver-cobalt veins occur at Gowganda. The future commercial importance of the several iron ranges of the area is very doubtful.

J. F. W.

Contributions to the Mineralogy of Black Lake Area, Quebec. By EUGENE POITEVIN and R. P. D. GRAHAM. Canadian Geological Survey, Mus. Bull. No. 27, 1918. Pp. 103, pls. 12, figs. 22.

A detailed study of the minerals of the chromite and asbestos pits in the Black Lake area, Megantic County, Quebec. This is a very productive area, in the serpentine belt of the eastern townships. The country rocks consist of a complex of igneous rocks, ranging from the most

basic to the most acidic in composition, and from late Cambrian to pre-Devonian in age. These igneous rocks probably take the form of thick laccoliths, and the different rock varieties are arranged in the order of decreasing basicity. In many cases erosion has removed the acidic members of the series. Serpentine itself is the least abundant rock of the area, but the most important economically.

Thirty-four mineral species are described from the area. In many cases their origin is given, especially the alumino-silicates rich in lime such as diopside, vesuvianite, and grossularite, which occur as dikes in the peridotite and are not the products of contact metamorphism. The CaO content for these minerals is thought to have been extracted by magmatic waters from the already consolidated portions of the igneous mass. Microscopic diamond crystals were found in the chromite, which is further evidence of the primary origin of chromite. Eleven new forms of diopside are recorded, with a number of illustrative drawings. Colerainite, $H_5Mg_2AlSiO_8$, is a new mineral species found in Coleraine Township, and its physical properties are described in detail with a number of chemical analyses. The mode of origin of the various varieties of serpentine is described with chemical analyses. Good views of the pits and microphotographs are given.

J. F. W.

Report on Braxton and Clay Counties. By RAY V. HENNER. West Virginia Geological Survey, 1917. Pp. 883, pls. 29, figs. 16.

A report on the mineral resources of the area with a discussion of its general geology. Aside from soils the principal wealth of the two counties is in the oil and gas pools, building-stone, and clay and shale for brick. The report is accompanied by topographic and geologic maps.

Part I considers briefly the physiography and history of the development of the region. The counties are in the central part of the state, on the eastern flank of the Appalachian geosyncline. Their present topography is that of a deeply dissected plateau.

Part II is an account of the general geology. The structure is simple, consisting of a gentle dip to the northwest, interrupted by gentle folds. The stratigraphic range is from the upper Devonian through the Paleozoic. Some Pleistocene river terrace deposits are present. A detailed description accompanied by sections is given for each formation present.

Part III discusses the mineral resources, the chief of which are oil and gas. Their development is of recent date. But few wells have been driven into the Chemung, and none below it, the present, known producing horizons being limited to the Pennsylvanian and Mississippian. Coal-mining operations, while on a large scale, are insignificant when compared to those of other counties of the state. The author estimates that the total available tonnage that may eventually be recovered is about 4,440,000,000. While there is not a single brick or pottery plant utilizing clays within the counties, there is an almost inexhaustible supply of raw materials as well as cheap fuel. Sandstone for road macadam and building purposes is abundant. In Clay County about half of the land is unfit for agricultural purposes, and it is suggested that this land be reforested.

Part IV consists of several paleontological contributions. W. A. Pierce presents some notes on the fossils of the Winefrede limestone and Uffington shale in which he notes the absence of a marine fauna. Professor E. C. Case describes the leg bone of a pareiasaurus-like reptile found in the Conemaugh series. I. C. White gives a few notes on the Conemaugh and Permian of the region, and comes to the conclusion that "not only the reptilian life, but also the plant and insect life of the Conemaugh series supports the conclusion that the beginning of red sediments in the Conemaugh marks the dawn of Permian time while there is nothing in the marine life of the epoch to contradict the same when properly interpreted.

Attached to the report is an appendix giving the elevations above mean tide for the area.

A. C. McF.

The Mackenzie River Basin. By Charles Camsell and Wyatt Malcolm. Canadian Geological Survey, Memoir 108, 1919. Pp. 154, pls. 14, fig. 1, and map.

This is a compilation of what is known concerning the geology of the Mackenzie River basin, which is about 1,350 miles long and 100 miles wide at the mouth of the river and 900 miles wide near the center, with a total area of about 682,000 square miles.

Parts of three chief physiographic provinces are included in this area and each one runs almost the whole length of the basin. They are the Laurentian Plateau on the east, the Great Central Plain of North America in the center, and the Cordilleran region on the west.

The outstanding characteristics of each of these three provinces is given along with a description of the Mackenzie River, including its lakes and larger branches.

Early pre-Cambrian rocks outcrop in the eastern part of the basin and consist of various schists, slates, limestones, and quartzites intruded by granites and gneisses. These are overlaid unconformably by sandstones, limestones, and basic flows and intrusives of late pre-Cambrian age. The Paleozoic is represented by a series of limestones, shales, and sandstones, not subdivided and of unknown age. A series of limestones and shales is classed as the Devonian, but the basal part is Upper Silurian in age, according to fossil evidence. Beds of gypsum are interbedded with these basal limestones, and the strata above the gypsum beds are fractured and folded, which is thought to be the result of expansion due to the alteration of beds of anhydrite to gypsum. The Mesezoic is represented by the Cretaceous sandstones and shales which occupy nearly the whole of the valleys of Athabasca and Peace rivers. Traced northward from Peace River these formations show three changes: a decrease in thickness, replacement of sandstone by shale, and a substitution of subaerial for marine conditions of deposition. In the sandstones in the basin of Athabasca River there are a number of workable seams of coal as well as extensive deposits of bituminous sands. A few small areas of Tertiary sands and clays overlie the Cretaceous with slight unconformity. Lignite seams occur in these beds.

Only the highest parts of the Rocky and Mackenzie mountains escaped Pleistocene glaciation. The ice from the Keewatin Glacier, which moved north, west, and south, entered this area, as well as ice from the mountains to the west. Glacial and lacustrine deposits are very extensive.

Descriptions are given of the bituminuous sands, with a discussion of their possible utilization, also notes on the various coal horizons, gypsum beds, salt springs, and clays of the area. Cobalt, gold, hematite, lead, zinc, and nickel are known to exist but very little is known as to their extent. A gas-bearing horizon in the Cretaceous has been known for twenty years, also petroleum from borings on the Peace River. During the last two years, active prospecting for gas and oil has been carried on with favorable results.

This is a valuable compilation and contribution to the geology of this little-known region. In the future it will serve as the starting-point for geologists working there. A good bibliography is given.

J. F. W.

Report on Berkeley, Morgan and Jefferson Counties. By G. P. Grimsley. West Virginia Geological Survey, 1916. Pp. 644, pls. 37, figs. 20.

This report covers an area of 768 square miles, comprising what is known as the "Eastern Panhandle," and is accompanied by topographic and geologic maps.

Part I is concerned with the physiography, climate and industrial development of the area.

Part II deals with the general geology. The structure is that typically developed in the Appalachians, consisting of parallel folds and faults. The deformation is not intense. Stratigraphically the rocks range from the Algonkian through the Carboniferous. Detailed and generalized sections, with local faunal lists, are given. Questions of correlation are discussed, and attention is given to the origin of the Catskill formation which is well developed in the region.

Part III is a discussion of the mineral resources, the more important of which include glass-sand, limestone clays, and road materials. There is included with each a discussion of the preparation and uses of the raw material.

As a result of the vast deposits of Cambro-Ordovician limestones, the limestone and lime industries are well developed. The Stones River formation contains limestone of great purity, and furnishes a high-grade fluxing material and a high-grade lime. An abundant supply of clay and shale for brick is available, the Martinsburg shale being of importance. Sandstone, quartzite, and limestone for road metal are found in abundance. Iron ore is of negligible importance but the region is admirably situated with respect to transportation, coal production and limestone fluxes, for the steel industry.

Attached to the report are two appendixes: (*a*) levels above the tide in the Eastern Panhandle region; (*b*) location of true meridian lines in the Eastern Panhandle region.

A. C. McF.

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VOLCANIC EARTHQUAKES

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Birmingham, England

Volcanic earthquakes, according to the late Professor Mercalli, are those which have their centers of maximum intensity under or close to the cones of active or semi-extinct volcanoes.[1] Professor Omori somewhat enlarges this definition. "A volcanic earthquake," he says, "may be defined as a seismic disturbance, which is due to the direct action of the volcanic force, or one whose origin lies under, or in the immediate vicinity of, a volcano, whether active, dormant or extinct."[2] Of the two definitions, the latter is the wider in its scope, for it includes the earthquakes which visit extinct volcanoes, such as the Alban Hills near Rome—earthquakes which differ in no important particular from those of an active volcano like Etna or of a dormant volcano such as M. Epomeo in the island of Ischia.

Mercalli follows up his definition of volcanic earthquakes by describing their important properties. The earthquakes, he says, (1) are felt many times in an area which is very restricted, although the shocks are violent; (2) they precede slightly, but sometimes accompany or follow, the eruptions of the neighboring volcano; and (3) they are repeatedly felt in the same area with similar characters

[1] G. Mercalli, *Vulcani e Fenomeni Vulcanici in Italia* (1883), p. 355.

[2] F. Omori, *Bull. Imp. Earthquake Inv. Com.*, Vol. VI (1912), p. 8.

so long as the volcano maintains its activity without notable change. Tectonic earthquakes differ in the following respects from volcanic earthquakes: (1) for the same intensity at the epicenter, they are felt over a much wider area; (2) though they sometimes occur in the immediate neighborhood of active volcanoes, they are as a rule quite independent of volcanic eruptions; and (3) great tectonic earthquakes seldom revisit the same district except at wide intervals of time. It will be seen later that these are not the only important differences between volcanic and tectonic earthquakes. For the present, however, it may be inferred that volcanic earthquakes as a rule originate in shallow foci which are confined to a limited region, while tectonic earthquakes spread from deeply seated foci that are subject to continual change.

Professor Omori notices that, under his definition, tectonic earthquakes may occasionally be included—earthquakes which disturb large areas[1] and are recorded by seismographs at very distant stations. For instance, about thirty hours before the eruption of the Usu-san (north Japan) on July 25, 1910, an earthquake occurred that was felt to a distance of 87 miles from the volcano and was recorded at a distance of 475 miles.[2] The beginning of the great eruption of the Sakura-jima (south Japan) on January 12, 1914, was followed after a few hours by an earthquake that damaged houses in Kagoshima and the surrounding country and was recorded in European observatories and probably in all parts of the world. Moreover, one month later, on February 13, another strong earthquake took place during an eruption of the Iwo-jima, a volcano belonging to the same chain as the Sakura-jima.[3] To these examples may be added the destructive Hawaiian earthquake of April 2, 1868, which originated in or near the southern part of Hawaii at nearly the same time as great eruptions of

[1] The occurrence of such earthquakes has long been recognized. For instance, G. P. Scrope, in his *Considerations on Volcanos* (1825), states that "those shocks . . . . which are felt to a considerable distance, are probably caused by new rents produced in the solid subjacent strata supporting or surrounding the mountain, and enter into that class of earthquakes which were discussed in a former chapter," that is, of tectonic earthquakes (p. 155).

[2] F. Omori, *Bull. Imp. Earthquake Inv. Com.*, Vol. V (1911), p. 15; Vol. VI (1912), p. 9.

[3] *Ibid.*, Vol. VIII (1914), p. 23; see also *Nature*, Vol. XCII (1914), pp. 716–17.

Kilauea and Mauna Loa, and was certainly felt in the island of Kauai, about 350 miles from the epicenter, or probably over an area of 375,000 square miles.[1]

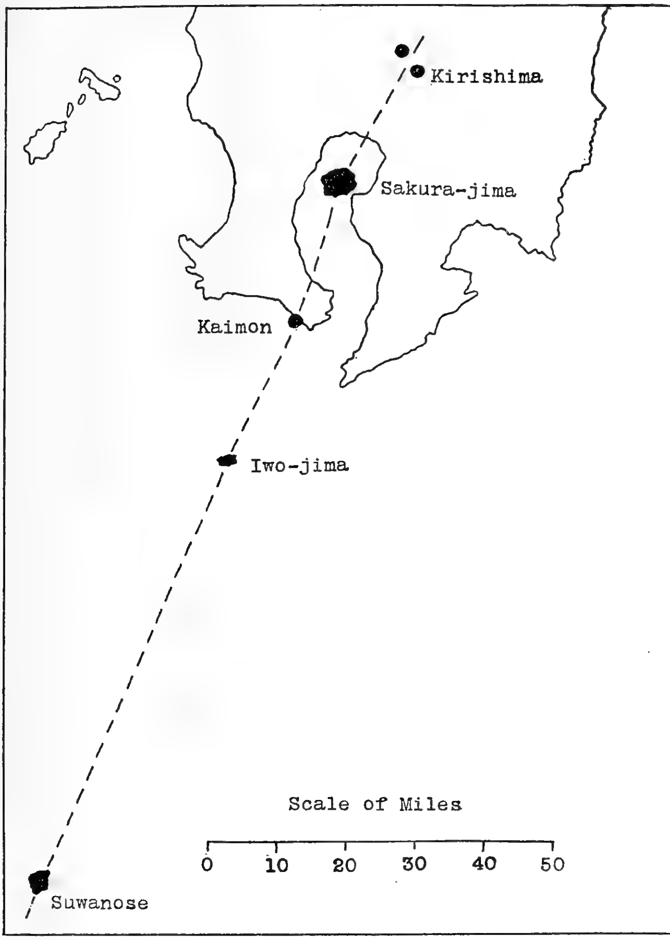

FIG. 1.—Map of volcanic chain in south Japan

It is of some consequence to decide whether such earthquakes should be regarded, as they would be under Professor Omori's definition, as volcanic earthquakes. The course of the volcanic

[1] See an admirable paper by H. O. Wood, "On the Earthquakes of 1868 in Hawaii," *Bull. Seis. Soc. of America*, Vol. IV (1914), pp. 169-203.

chain of southern Japan referred to in the last paragraph is shown in the accompanying sketch map (Fig. 1), and it is interesting to notice the progressive awakening from north to south of the volcanic foci along its course. On November 18, 1913, the Kirishima-yama broke out in strong eruption, which lasted into the following year. The Sakura-jima followed on January 12, 1914; and, about a month later, the Iwo-jima. When three volcanoes, situated as these are and all of infrequent activity, break into eruption so nearly together, and when two of the eruptions are accompanied by strong and deeply seated earthquakes, it would seem natural to ascribe both phenomena to a common cause—the earthquakes directly, and the eruptions indirectly—to the stress accumulation along the whole volcanic chain.[1]

It seems to me, then, that tectonic earthquakes which originate in the immediate neighborhood of volcanoes and even concurrently with eruptions of the same, are not directly of volcanic origin, and should not, by reason of such proximity, be regarded as volcanic earthquakes. I venture, therefore, to suggest that the definitions of such earthquakes given by the two distinguished seismologists referred to should be somewhat modified and should be replaced by the following: A volcanic earthquake is an earthquake directly due to the operations which result or tend to result in a volcanic eruption or is due to relative movements, by whatever cause they may be produced, along fractures of the volcanic mass, whether the volcano itself is active, dormant, or extinct.

Adopting this as the definition of a volcanic earthquake, I propose in the first section of this paper to describe the earthquakes connected with a few typical volcanoes, namely, certain Japanese volcanoes (the Usu-san, the Asama-yama, and the Sakura-jima) and Etna, as examples of active volcanoes, and M. Epomeo in Ischia and the Alban Hills near Rome as examples of dormant and extinct volcanoes respectively. In the second section is given a summary of the characteristic phenomena of volcanic earthquakes; and in the third and last section the modes of origin of volcanic earthquakes will be considered.

[1] F. Omori, *Bull. Imp. Earthquake Inv. Com.*, Vol. VII (1914), pp. 23–24.

I. DESCRIPTION OF SOME VOLCANIC EARTHQUAKES

1. JAPANESE VOLCANOES

Usu-san and Sakura-jima.—The Usu-san is situated near the southwest end of Hokkaido, the northern island of Japan, and the Sakura-jima in the Bay of Kagoshima on the south coast of Kyushu, the southern island. The last eruption of the Usu-san began at 10 P.M. on July 25, 1910, and that of the Sakura-jima at 10 A.M. on January 12, 1914.

Both eruptions were preceded by a large number of earthquakes. Those of the Usu-san outburst began on July 21. At Nishi-Monbets, about 5 miles from the center of the volcano, 25 shocks were felt on July 22, 110 on July 23, 351 on July 24, and 165 on July 25. Once the eruption had begun, the seismic frequency decreased. At Sapporo, about 44 miles from the volcano, the earthquakes were registered by a horizontal pendulum seismograph, the numbers being 1 (at 4:18 P.M.) on July 21, 3 on July 22, 23 on July 23, 76 on July 24, 84 on July 25; after the eruption there was a rapid decline in frequency, the numbers being 26 on July 26, 15 on July 27, 5 on July 28, 6 on July 29, and 1 on July 30. The eruption continued until the end of the year.[1]

At Kagoshima, which is about 6 miles from the center of the Sakura-jima, the first earthquake was felt at 3:41 A.M. on January 11, 1914. At the Kagoshima Observatory the earthquakes were recorded by a Gray-Milne seismograph, the average hourly frequency being 4.1 from 3 to 11 A.M. on January 11, 12.4 from 11 A.M. to 8 P.M., and 19.5 from 8 P.M. on January 11 to 10 A.M. on January 12. The greatest hourly numbers were 28 at 8–9 P.M. on January 11 and 27 at 3–4 A.M. on January 12. The total number of shocks registered from 3 A.M. on January 11 to 10 A.M. on January 12 was 418. After the first eruption at the last-mentioned hour, there was a marked decline in earthquake frequency, the numbers being: 10–11 A.M., 17; 11–noon, 11; noon–1 P.M., 6; 1–2 P.M., 3; 2–3 P.M., 5; 3–4 P.M., 2; 4–5 P.M., 2; 5–6 P.M., 2. At 6:30 P.M., the seismograph was injured by the strong tectonic earthquake referred to above.[2]

[1] *Ibid.*, Vol. V (1911), pp. 8–17.

[2] *Ibid.*, Vol. VIII (1914), pp. 9–14, 22–27.

Thus, both eruptions were preceded by a marked increase in seismic frequency, followed by a marked decrease, the maxima occurring 24 hours before the first outburst of the Usu-san and about 13 and 6 hours before that of the Sakura-jima.[1]

Observations were made with a portable horizontal tromometer erected by Professor Omori at Nishi-Monbets (5 miles from the Usu-san) from July 30 to August 6, and at the West-Kohan School (Sobets) at the foot of the East Maru-yama from August 6 to 10. At this place, which is close to the nearest craterlet, series of well-defined, small, quick, unfelt vibrations, called micro-tremors by Professor Omori, were registered, which were entirely absent from the records at Nishi-Monbets. The mean range of motion was in every case less than one-tenth of a millimeter; but the principal periods of the tremors (.53, 1.08, 1.59, 2.14 seconds) were practically identical with those of earthquake vibrations recorded at Nishi-Monbets (.53, 1.01, 1.58, 2.43 seconds). It would seem, then, that the micro-tremors are in reality true earthquake vibrations, but so weak that they cannot be recorded more than a few miles from the origin. Professor Omori notices that the shortest of the foregoing periods is approximately one-half, one-third, and one-quarter of the other periods. He also shows that moderate explosions from even the nearest craterlet were not as a rule accompanied by marked micro-tremors; whereas violent explosions from that and other craterlets were usually accompanied, and often preceded by several minutes, by well-pronounced micro-tremors. The tremors, however, were not confined to the epochs of explosions. They sometimes occurred when the smoke ejections from the different craterlets were insignificant and even when they had completely ceased. At such times, as Professor Omori suggests, eruptions were perhaps prevented by the temporary stoppage of the craterlets.[2]

Asama-yama.—The Asama-yama, one of the greatest of Japanese volcanoes, rises from the plateau of the central island of

[1] Professor Omori's observations give precision to a fact which has long been known. "It is," says G. P. Scrope in his *Considerations on Volcanos* (1825), "a remark common to the observations made on almost all volcanic eruptions, that local earthquakes always precede the emission of lava currents, and cease while the lava is flowing, to recommence when it has stopped" (p. 155).

[2] *Op. cit.*, Vol. V (1911), pp. 31–38.

Japan to a height of 8,140 feet above the sea. After a prolonged period of rest, it has been subject for about six years to a series of strong explosions, the first of which occurred on February 13, 1908.

Observations with horizontal pendulum seismographs were instituted by Professor Omori at two stations on the southwest flank of the volcano, Yuno-taira and Ashino-taira, at heights of 6,306 and 4,422 feet above the sea. Owing to weather conditions, the observations at the former station were confined to the summer months.

The seismograms obtained at these places showed that there were two distinct types of earthquakes, some being independent of any outburst of the volcano, while others were invariably the results of explosions.

The two types of earthquakes are characterized by several marked differences, of which the following are the more important:

a) The shocks without explosions consisted only of minute quick vibrations, while those with explosions consisted of slow movements (of as much as 2.6 and 5.3 seconds' period), on which after a few seconds quick vibrations were superposed.

b) The earthquakes without explosions were distinctly stronger than the others, probably because, as Professor Omori suggests, a great part of the energy of the explosions is expended in the projection of rock fragments and débris. Of 1,485 earthquakes without explosions, 21 per cent were sensible at Yuno-taira; of 8,847 earthquakes with explosions, only 0.3 per cent were sensible. Moreover, the strong earthquake of May 26, 1908, which did not accompany an explosion and which evidently originated in the volcano itself, was felt over an area of 2,400 square miles.[1]

c) The earthquakes without explosions were of shorter duration than those with explosions, the averages for the former being 16.7 seconds in 1911 and 15.0 seconds in 1912, and for the latter 33.2 seconds in 1911 and 32.7 seconds in 1912.

d) The two types of earthquakes alternate in frequency. During the two years, 1911–12, the maxima of the earthquakes

[1] It may be added that the Usu-san explosions of 1910 were preceded by many sensible earthquakes, a few of them strong; but, as a rule, they were unaccompanied by sensible shocks.

without explosions occurred at 3–4 A.M., 8–9 A.M., 1–2 P.M., and 8–9 P.M., while the minima of the earthquakes with explosions occurred at 3–4 A.M., 8–9 A.M., 2–3 P.M., and 9–10 P.M. The minima of the former occurred at 6–7 A.M., 11–12 A.M., 6–7 P.M., and 11–12 P.M., while the maxima of the latter occurred at 5–6 A.M., noon–1 P.M., 4–5 P.M., and 1–2 A.M. In August, 1911, the earthquakes with eruptions attained their maximum frequency (205) for the year, and the earthquakes without eruptions their minimum frequency (38); in October, 1912, there was a maximum of earthquakes with eruptions (626) and a minimum of earthquakes without eruptions (10). A similar relation is shown in the variation of frequency from one year to another, as will be seen in the following table:[1]

	Earthquakes without Eruptions			Earthquakes with Eruptions		
	Sensible	Unfelt	Total	Strong Explosions	Small Outbursts	Total
1911........	57	321	378	0	577	577
1912........	124	563	687	0	1111	1111
1913........	6	28	34	25	7101	7126
1914........	11	37	48	1	30	31
1915........	44	65	109	0	0	0
1916........	64	165	229	0	2	2

2. ETNA

Etna rises from a nearly circular base, measuring 428 square miles, to a height of 10,870 feet. Earthquakes occur on all sides of the mountain, and, for many years past, they have been especially frequent beneath its southeastern flank. In the present section, I propose to describe a few typical earthquakes on this flank, and to refer very briefly to the distribution of the epicenters within the area covered by the volcano.

The earthquakes belong as a rule to one of two classes, according as the greater axes of their meizoseismal areas are directed nearly along or perpendicular to the radii from the central crater. They may therefore be called *radial* and *perimetric* earthquakes, respectively.

[1] After May 5, 1914, there was no strong outburst of the Asama-yama.

As examples of radial earthquakes on the southeastern flank, I have selected three which have been studied in perhaps greater detail than others, namely, the Fondo Macchia earthquake of July 19, 1865, the Fleri earthquake of August 8, 1894, and the Linera earthquake of May 8, 1914. The Fondo Macchia earthquake of October 15, 1911, is given as an example of a perimetric earthquake.

Radial earthquakes.—On January 30, 1865, a great eruption took place on the east-northeast flank of Etna, which lasted nearly twelve weeks. Eighty-eight days after its close, on July 19, Fondo Macchia was completely destroyed by an earthquake, and the neighboring villages of Baglio, Rondinello, Scaronazzi, and S. Venerina were seriously damaged, though not ruined. The meizoseismal area was a narrow band directed W.NW. and E.SE., a little over 4 miles in length and about $\frac{2}{3}$ mile wide. Outside this band, the shock was disastrous within an area 5 miles long and $1\frac{1}{4}$ miles wide, represented by the curve *A* on the sketch map (Fig. 2). The intensity of the shock diminished rapidly toward the north, and more slowly toward the south, but at no place more than 12 miles from the epicenter was the shock felt. The area disturbed cannot therefore have exceeded 113, and was probably less than 100, square miles. A number of after-shocks followed this earthquake, 24 being felt in July (3 of them very strong at

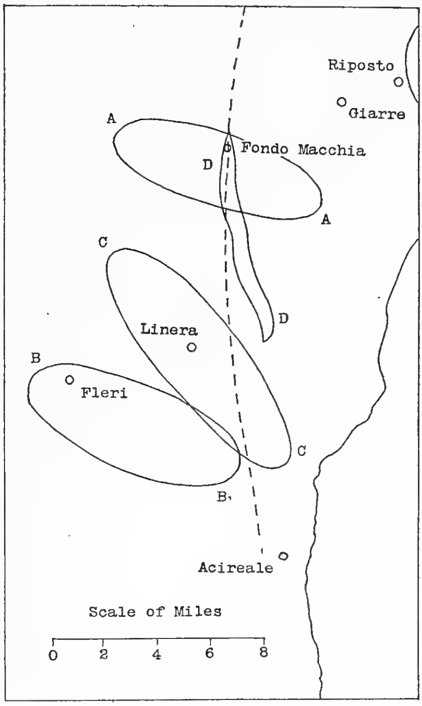

FIG. 2.—Map of earthquakes on southeast flank of Etna.

Fondo Macchia), and several others, more or less slight, in the following month, the series ending on August 23.[1]

The earthquake of August 8, 1894, differed in two respects from many Etnean earthquakes. It occurred while the volcano was in a state of only moderate activity, and it disturbed an area unusually large for the district, although small for a shock of such intensity. It was preceded by a violent shock at 1:58 P.M. on August 7, strongest at Fleri, Zerbate, etc., and felt generally at Catania and Zafferana, and by a few persons at Nicolosi and Trecastagni. Three other shocks followed in the meizoseismal area before the occurrence of the principal shock of the series at 6:16 A.M. on August 8. By this shock, the villages of Fleri, Pisano, Zerbate, etc., were ruined. The meizoseismal area, about 4 miles long from northwest to southeast and 2 miles wide, is represented by the curve *B* in Figure 2. From this area, the intensity of the shock diminished outward rather, but not very, rapidly, the disturbed area including the whole base of Etna and probably on the whole as much as 800 square miles. Within or near the epicentral district, at least 15 after-shocks were felt before the end of the month.[2]

Few, if any, Etnean earthquakes have thrown so much light on the nature and origin of volcanic earthquakes as the remarkable series which culminated in the Linera earthquake of May 8, 1914. They have been studied by Professor G. Platania in one of the most valuable memoirs that we possess on volcanic earthquakes.[3] In all respects except in the shallowness of the foci, they resemble true tectonic earthquakes. They were preceded and followed by long series of accessory shocks, and, along the axis of the meizoseismal area, there were displacements of a pre-existing fault.

Omitting instrumental shocks, the whole series contained 55 earthquakes, 21 being fore-shocks from April 28 to May 7, and 33 after-shocks from May 8 to June 4. Five of the fore-shocks and

[1] A. Riccò, *Boll. Soc. Sism. Ital.*, Vol. XVI (1912), pp. 27–31; M. Baratta, *Boll. Soc. Geogr. Ital.* (1894), pp. 12–13, and *I Terremoti d'Italia*, pp. 442–43.

[2] M. Baratta, *Boll. Soc. Geogr. Ital.* (1894), pp. 6–9, 23.

[3] "Sul periodo sismico del maggio 1914 nella regione orientale dell'Etna," *Pubbl. dell'Ist. di Geogr. Fis. e Vulcan. della R. Univ. di Catania*, No. 5, 1915.

seven of the after-shocks were strong. The broken line on the map (Fig. 3) includes all the places in which houses were damaged by the different shocks of the series. The smaller curves represent the meizoseismal areas of the more important shocks. Thus, the curves 1 and 2 indicate the meizoseismal areas of the double earthquake of May 7 at 6:35 P.M., houses being damaged at Piano d'Api (No. 1) and at Pennisi and Fiandaca (No. 2). At 10 P.M. on the same day, another strong earthquake caused damage at Fossalacqua (No. 3) along a fracture which may be a continuation of that at Fiandaca (No. 2). The principal earthquake occurred at 7:2 P.M. on May 8, and was ruinous within the area bounded by the ellipse (No. 4), the zone of complete ruin being much less wide. With this earthquake also there was a zone of serious damage to houses at Dagala (No. 5), separated from the principal zone by a wide tract of little or no damage to property. Lastly, the curve No. 6 shows the boundary of slight damage near Viagrande wrought by the after-shock of May 26.

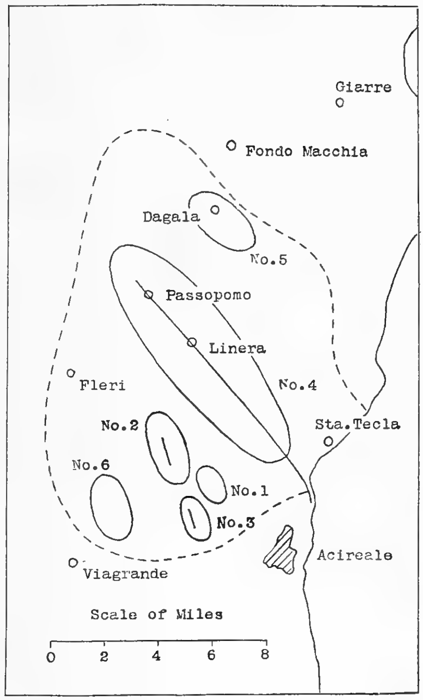

FIG. 3.—Map of Linera earthquakes of 1914

The meizoseismal area of the principal earthquake (bounded by the curve *C* in Fig. 2 and No. 4 in Fig. 3), within which the ruin was practically complete, is very elongated, being about $4\frac{1}{4}$ miles long, with a maximum width near Linera of $1\frac{1}{4}$ miles. In this zone, not only are the houses completely razed to the ground, but the ground itself is crushed. Along the axis of this zone, there runs a slightly sinuous fracture with, in parts, a secondary nearly parallel

fracture separated by a distance of from 2 to 50 or more yards. The most destructive effects of the earthquake are concentrated along this fracture, starting from Passopomo, through Linera, to beyond Mortara. Here the railway line was seriously damaged, the rails being displaced and contorted. The fracture can be followed as far as the seacoast, bending slightly to the south in the neighborhood of Sta. Tecla. Almost throughout the fractured zone there is a change of level, in some places of only an inch or two in others of 15 or 16 inches, and in one, near the seacoast, of more than 3 feet, the ground of the southwest side being left at a higher level than that on the northwest side.[1] The fracture is by no means a recent one, for it has been known since 1879, and Professor Platania states that displacements have occurred along it in previous earthquakes. Nor is it the only fracture along which movements have taken place during this series of earthquakes. There is a second at Fiandaca (No. 2), which was observed during the earthquakes of 1894 and 1907, and which probably corresponds with another at Fossalacqua (No. 3). Near this fracture occurred the greatest damage wrought by the two strong fore-shocks of May 7.

During the interval covered by this series of earthquakes there was a marked increase in the activity of Etna, though the different shocks were not coincident with the volcanic explosions.

Perimetric earthquakes.—On September 10, 1911, a great eruption of Etna began, and, notwithstanding the extraordinary energy of the early phenomena, ceased after only thirteen days of activity. Three weeks later occurred the destructive Fondo Macchia earthquake of October 15, 1911. This earthquake was, however, preceded by at least 10 fore-shocks, the first 5 of which occurred on September 30 at Pisano (intensity 7, Mercalli scale), October 9 (2 shocks) at Piano d'Api (intensity 5), and October 14 at S. Venerina (intensity 6). The principal shock (intensity 10) occurred at 9:52 A.M. on October 15, and was followed by 4 slight after-shocks on October 24 and 1 on November 9.

[1] There is also some evidence of horizontal displacement. A high wall in the upper part of the fractured zone is curved and shifted toward the southeast.

The isoseismal lines of intensities 10, 8, 6, 4, and 2 (Mercalli scale) are shown in Figure 4, the line of intensity 10 being also represented by the curve *D* in Figure 2. The meizoseismal area (bounded by the isoseismal 8) is a slightly sinuous band, running from N.NW. to S.SE., 4 miles long, about $\frac{1}{3}$ mile wide, and $1\frac{1}{4}$ square miles in area. Within it is a band, including Fondo Macchia, in which the destruction of buildings was complete, this band being about 3 miles long and $\frac{1}{4}$ mile wide. Notwithstanding the great intensity within this band, the shock was not felt at places 6 miles to the west, the area within the isoseismal 4 contained not more than about 70 square miles, and the whole district shaken only about 230 square miles. The mean duration of the shock was about 8 seconds.

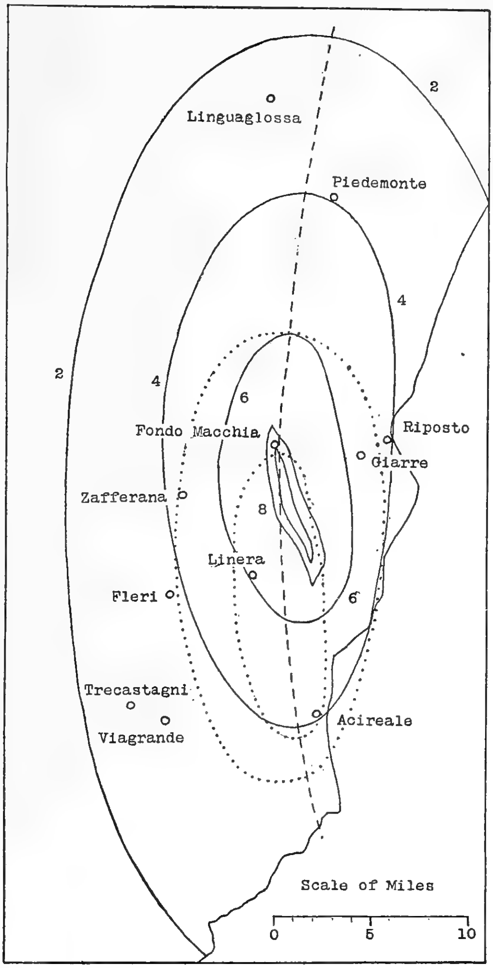

FIG. 4.—Map of Fondo Macchia earthquake of October 15, 1911.

During the eighteen years (1893–1911) preceding this earthquake, 27 shocks more or less strong originated within the elliptical

area represented by the outer dotted line in Figure 4. Six of these earthquakes were of ruinous strength, with their epicenters at Fondo Macchi, Zerbate, S. Leonardello, and Aci Platania, all included within the area represented by the smaller dotted ellipse in Figure 4. The two ellipses have their longer axes coinciding with those of the isoseismal lines of the Fondo Macchia earthquake of 1911. Close to these axes and nearly parallel to them, runs a fault known as the *Timpa della Scala* and represented by the broken line on the map. The western limb at the south end has undergone, and is still undergoing, elevation; while the eastern limb at the north end is subsiding.[1]

Distribution of Etnean earthquakes.—A few of the Etnean earthquakes disturb the whole area of the volcano; but the majority are strongly felt in one or a few of the villages scattered over the mountain sides. In such cases the villages affected cannot be far distant from the epicentral areas, and Dr. M. Baratta, in his great history of Italian earthquakes,[2] has thus found it possible to distinguish the principal seismic zones of this volcanic region. These are twelve in number, the places which give their names to the different zones being shown in Figure 5, in which the dotted line marks out the base of the volcano: (1) Linguaglossa, to the NE.; (2) Randazzo, to the N.; (3) Adernò-Bronte-Maletto, to the W.; (4) Santa Maria di Licodia, to the S.W.; (5) Paternò, to the S.SW.; (6) Belpasso, to the S.; (7) Nicolosi, to the S.; (8) Trecastagni, to the S. SE.; (9) Acireale, to the SE.; (10) Zafferana-Pisano-S. Venerina, to the E.SE.; (11) Macchia Region, to the E.; and (12) Giarre-Riposto, to the E.

Of these zones, the most important at present are those of Santa Maria di Licodia, Nicolosi, Trecastagni, Acireale, Zafferana-Pisano-S. Venerina, and the Macchia Region. The Fleri earthquake of 1894, described above, is included in the Acireale zone, the Linera earthquake of 1914 in the Zafferana zone, and the Fondo Macchia earthquakes of 1865 and 1911 in the Macchia zone. The Nicolosi zone gives rise to many and violent earthquakes, and it is remarkable that some of the greatest are quite

[1] A. Riccò, *Boll. Soc. Sis. Ital.*, Vol. XIX (1912), pp. 9-38.

[2] *I Terremoti d'Italia*, pp. 829-33.

local. For instance, the earthquake of May 11, 1901, damaged many houses in Nicolosi (intensity 8), and yet was only just perceptible at a distance of 3 or 4 miles.[1] The earthquake of May 14, 1898, is typical of the S. Maria di Licodia zone; nearly all the

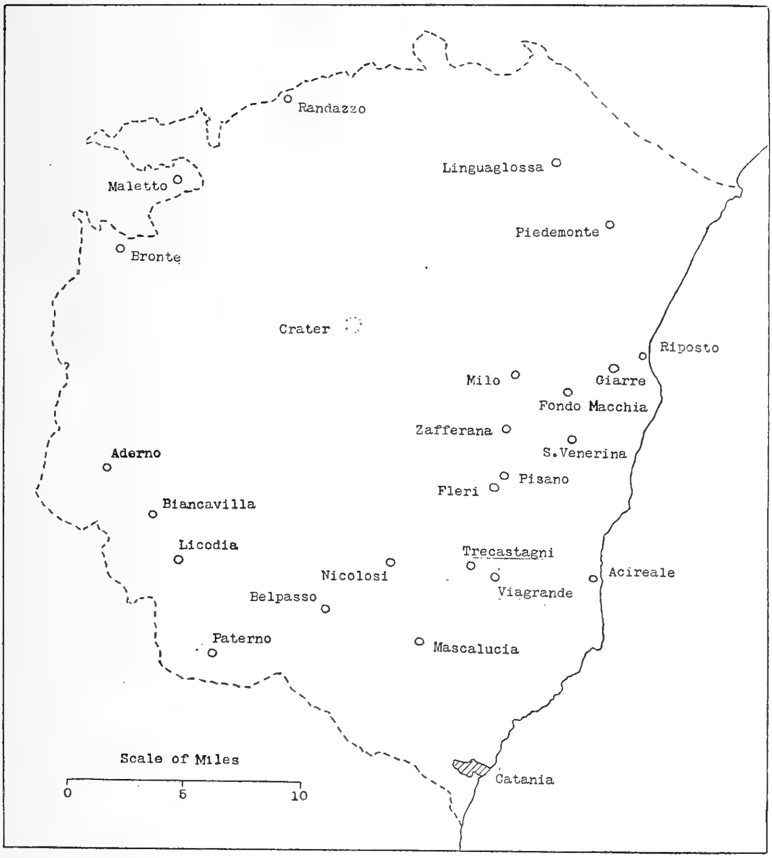

FIG. 5.—Map of Etna and seismic districts

houses in S. Maria di Licodia were damaged, and slight injury occurred in other villages on the southwest slope from Belpasso to Adernò, the shock being thus rather more widely felt than the Nicolosi earthquake.[2]

[1] S. Arcidiacono, *Boll. dell' Accad. Gioenia di Sci. Nat. in Catania*, Fasc. 70 (1901).

[2] A. Riccò, *op. cit.*, Fasc. 53–54 (1898).

For many years past the most active zones have been those between the eastern and southern radii. But the seismic foci migrate from one zone to another without (at present) any apparent law. Take, for instance, the earthquakes of one year only, the year 1903.[1] On January 30 there was a shock of intensity 5 at Giarre, Zafferana, Milo, Viagrande, Linguaglossa, and Randazzo (the epicenter being perhaps in zone 12); on March 8, one of intensity 4 between Nicolosi and Mascalucia, but hardly felt at either of these places although they are only 4 miles apart (zone 7); on March 11, one of the same intensity at S. Venerina (zone 10); on March 24, a very slight shock at Belpasso (zone 6); on April 3, a shock of intensity 5, and on April 4 and 5 very slight shocks at S. Venerina (zone 10); on April 7, a shock of intensity 5 at Linguaglossa and Milo (zone 1); on April 13, one of intensity 4 at Biancavilla (zone 4?); on April 19, a slight shock at Paternò (zone 5); on May 26, a very strong shock with slight damage at Trecastagni (zone 8), and only feebly felt at Viagrande, less than a mile away; on June 1, 14 shocks (2 of them of intensity 6) at Pedara and Viagrande; on June 3, 3 shocks at Viagrande; on June 5, 1 shock, and on June 7, 2 shocks, at Trecastagni and Viagrande; on June 11, a slight shock at Trecastagni; and on June 16, 2 very slight shocks at Viagrande (all 23 shocks in zone 8); on July 21, a very slight shock at Belpasso (zone 6); on July 30, a sensible shock at Biancavilla (zone 4?); on August 6, 4 shocks of intensity 4 at Nicolosi, Trecastagni, Zafferana, Milo, and Biancavilla (zone 7?); and, lastly, on November 20, a strong earthquake of intensity 6 at Viagrande, Zafferana, Milo, S. Venerina, Acireale, and Linguaglossa (zone 10). Thus, seven out of the twelve Etnean zones were probably in action in this one year.

3. MONTE EPOMEO (ISCHIA)

The island of Ischia, 6 miles long from east to west and 5 miles in width, is separated from the Italian coast by a distance of only 6 miles. A large part of the island is occupied by the old crater of Monte Epomeo, a volcano which must be regarded as dormant rather than extinct, for the last eruption occurred in

[1] S. Arcidiacono, *Bull. dell' Accad. Gioenia di Sci. Nat. in Catania*, Fasc. 79 (1903).

1302, and the one before that a thousand years earlier. These eruptions, as well as their more recent predecessors, took place from new vents chiefly on its eastern and northeastern flanks. All of them began suddenly, were accompanied by violent earthquakes, and were preceded by long intervals of repose.

Such an interval, one of more than four centuries, followed the last eruption, and during that time there were no earthquakes

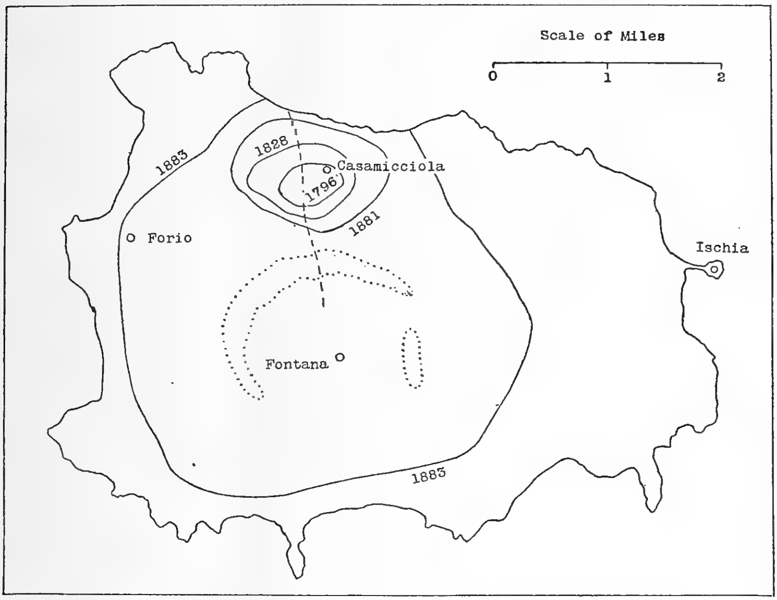

FIG. 6.—Map of Ischian earthquakes

of any consequence in the island. Then, during one night—July 28–29, 1762—as many as 62 earthquakes were felt at Casamicciola, some of them strong enough to damage buildings. On March 18, 1796, a severe earthquake occurred, which caused destruction of buildings in Casamicciola alone; on February 2, 1828, one still stronger, which again wrought destruction in that town; others of less consequence, but still strong, occurred on March 6, 1841, and August 15–16, 1867; the last of all being the destructive earthquakes of March 4, 1881, and July 28, 1883.

In the accompanying map (Fig. 6) the dotted lines represent the boundaries of those portions of the central crater of Epomeo

which are still standing, and the continuous lines the boundaries of the areas within which buildings suffered marked damage by the earthquakes of 1796, 1828, 1881, and 1883. It will be noticed that these areas show a progressive increase in size. The broken line, slightly curved, shows the position of the radial fracture with which the earthquakes were probably connected; and it is important to notice that the chief damage wrought by the earthquakes of 1881 and 1883 was concentrated along this line near Casamicciola.

These earthquakes were studied in great detail by the late Dr. H. J. Johnston-Lavis. The isoseismal bounding the area of complete destruction in 1881 includes only $\frac{1}{2}$ square mile, the area of partial though serious destruction about 2 square miles, and that within which buildings were slightly damaged about 5 square miles. Still farther, the shock continued to diminish rapidly in intensity. It was felt in the neighboring island of Procida, and by some persons, though very slightly, on the Italian coast, the total area being probably less than 300 square miles. From the inclination of the fissures made in buildings, Dr. Johnston-Lavis estimated the mean depth of the focus at about $\frac{1}{3}$ mile. The earthquake was followed by a few slight after-shocks on March 7 (2 shocks), 11–12, 15–16, 17–18, and 27(?), April 5 and 6, and July 18.

The earthquake of July 28, 1883, was much stronger, and resulted in great loss of life at Casamicciola. It was preceded at that place by a slight shock on July 24, and by an earth-sound a quarter of an hour beforehand. The areas of complete, partial destruction and slight damage were respectively 3, 11, and 30 square miles. In this earthquake, also, the intensity diminished very rapidly outward. The shock was felt in Italy near the coast and by a few persons in Naples, which is 20 miles from Casamicciola. The disturbed area can hardly therefore have contained more than 1,250 square miles. The mean depth of the focus was again found to be about $\frac{1}{3}$ mile. The shock was remarkable for the entire absence of preliminary sound or tremor and for its great initial strength, a great part of the ruin being caused during the first few seconds. Between July 28 and August 3, 21 slight

after-shocks were felt in Casamicciola. This, however, was not the only center in action, for several were recorded at the town of Ischia only.[1]

4. ALBAN HILLS

About 15 miles to the southeast of Rome lies a group of extinct volcanoes containing several circular lakes which evidently occupy old craters of the system. The principal hill, Monte Cavo, of which the other and smaller hills may be regarded as lateral cones, rises to a height of nearly 3,000 feet. There is no certain evidence of activity during historic times.

The earthquakes of this district are less frequent and destructive than those of the much larger Etnean region, but in other respects they resemble them closely, and especially in the constant migration of activity from one part of the volcano to another.

One of the most recent in the district is that of February 21, 1906. It was preceded by six very slight fore-shocks, one on February 20, at Ariccia, the others on February 21, at Albano. The principal shock occurred at 9:49 P.M. on February 21, and, from that time until February 23, there were 6 very slight after-shocks. The isoseismal lines of the principal earthquake, which was of intensity 5–6 (Mercalli scale), are represented approximately in Figure 7. They are rather close together, indicating a rapid decline in the intensity of the shock from the epicenter, though a decline less rapid than in other earthquakes of the same district or of the Etnean region. The mean duration of the shock was nearly 4 seconds.[2]

Dr. M. Baratta defines the following seismic zones among the Alban Hills (Fig. 7): (1) Colonna; (2) Montecompatri; (3) Monte Porzio; (4) Frascati; (5) Rocca di Papa-Monte Cavo; (6) Albano; (7) Ariccia; (8) Nemi; and (9) Velletri.[3] To these districts should perhaps be added: (10) Genzano-Civitalavinia.

[1] G. Mercalli, *L'isola d'Ischia ed il terremoto del 28 luglio 1883* (Milan, 1884); H. J. Johnston-Lavis, *Monograph of the Earthquakes of Ischia* (1885). A summary of these and other memoirs on the Ischian earthquakes is given in my *Study of Recent Earthquakes*, pp. 45–73.

[2] G. Agamennone, *Boll. Soc. Sis. Ital.*, Vol. XXI (1918), pp. 47–101.

[3] *Terremoti d'Italia*, pp. 778–80.

Of these zones, Nos. 1–4 lie on the northern half, and Nos. 6–10 on the southern half, of the volcano.

As in the Etnean region, there is at present no apparent law in the transference of the epicenters from one zone to another. The nature of that transference may be shown by a brief reference to the more important earthquakes of the last forty years. On September 2, 1883, an earthquake of intensity 7 (Mercalli scale)

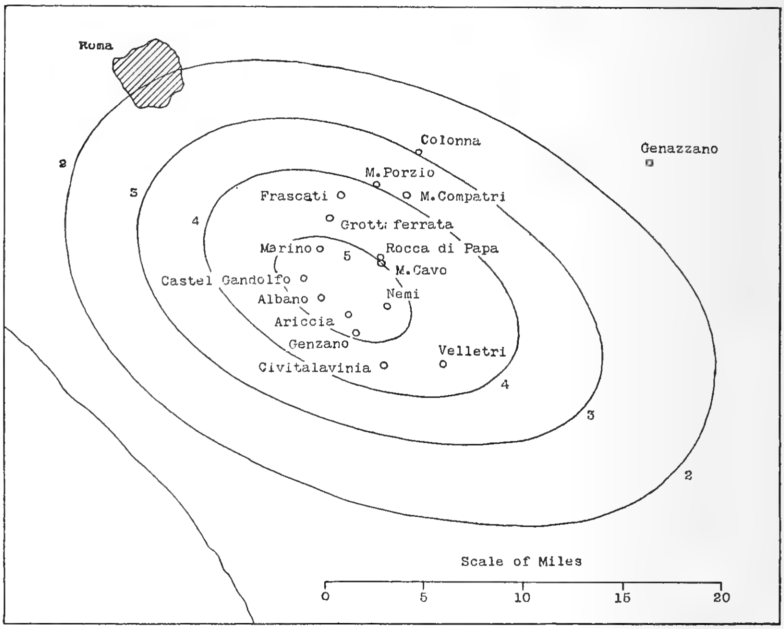

FIG. 7.—Map of Albano earthquake of February 21, 1906

caused slight damage at Frascati, Grottaferrata, Rocca di Papa, and Monte Cavo (zone 4), the axis of the area of maximum intensity being directed southeast or along a radius from the summit of Monte Cavo; on January 17, 1886, there was a very strong shock at Ariccia (zone 7); on January 22, 1892, a shock with its epicenter between Genzano and Civitalavinia (zone 10); on May 8, 1897, a strong shock without damage at Colonna, the axis of the area of maximum intensity being directed southwest or along a radius (zone 1); on November 6, 1897, a shock at Civitalavinia

(zone 10), followed a week later (on November 13) by one with its epicenter near Velletri (zone 9); on July 19, 1899, the strongest shock of the series (intensity 8), causing some damage at Frascati (zone 4) and rather less at Marino, the mean duration of the shock $3\frac{1}{2}$ seconds, registered at Lubiano (298 miles to the north) and Catania (318 miles to the south); on October 12, 1902, a shock of intensity 6 at Genzano (zone 10); on February 21, 1906, one of intensity 5–6 at Albano (zone 6); and, lastly, on October 6, 1909, a shock at Frascati (zone 4). Thus, of these ten earthquakes, four originated on the north side, and six on the south side, of the volcano; while six of the ten seismic zones were in action.

II. CHARACTERISTICS OF VOLCANIC EARTHQUAKES

a) While volcanic earthquakes often occur in close connection with eruptions—either shortly before, during, or shortly after them—they seldom coincide with an outburst, except in the case of the weak tremors with which the Japanese records have made us familiar. Nor do the almost contemporaneous earthquakes and eruptions always occur in the same part of the volcano. The Fondo Macchia earthquakes of 1865 and 1911 were felt in a district to the east of the central crater of Etna, while they followed closely on eruptions from the northeastern flank. Some earthquakes, again, occur in active volcanic regions without the accompaniment of any change in volcanic activity; others in regions in which the volcanoes have been extinct for many centuries.

b) In volcanic earthquakes the intensity of the shock is often great near the center of an extremely small disturbed area. The relation between the intensity and the area is by no means constant. On the one hand, we may have an earthquake like that of Nicolosi in 1901, destroying houses in a minute epicentral area and yet imperceptible at a distance of more than 4 miles, or one like the Ischian earthquake of 1883, leveling every building within an area of 3 square miles, and yet felt within an area of not more than 1,250 square miles. On the other hand, we may have an earthquake like the Albano earthquake of 1906, not strong enough to damage houses at the epicenter and yet felt over an area of 535 square miles.

c) The sound and tremor, with which tectonic earthquakes usually begin, are either absent from volcanic earthquakes, or else of duration so short as to be almost imperceptible.

d) The duration of the shock in volcanic earthquakes is usually short, being very often less than 5 seconds.

e) The length of the focus is inconsiderable when compared with that of tectonic earthquakes. Even in the strongest of volcanic earthquakes, it rarely exceeds 4 or 5 miles, whereas the average length of the focus in British earthquakes (omitting the weakest) is about 12 or 13 miles. This conclusion is also supported by the brief duration of the shock.

f) Some, but not all, volcanic earthquakes are preceded and followed by a rather large number of accessory shocks. In this they resemble tectonic earthquakes. But the after-shocks of volcanic earthquakes are distinguished by the short period of their action. For instance, the series of after-shocks in the Albano earthquake of 1906 lasted for 3 days, in the Ischian earthquake of 1883, for 7 days, in the Fleri earthquake of 1894, for about 3 weeks, in the Linera earthquake of 1914, for nearly 4 weeks, and in the Fondo Macchia earthquake of 1865, for 5 weeks.

(*g*) The after-shocks of volcanic earthquakes are practically confined to the epicentral area. They point to little, if any, tendency toward any extension of the original focus.

h) In a volcanic system such as that of Etna or the Alban Hills earthquakes occur in many different zones, and seismic activity is subject to frequent and sudden migrations from one zone to another. Nevertheless, in any given zone there is often a certain fixity in the epicenters of successive earthquakes, as in those of Fondo Macchia (Etna) and Casamicciola (Ischia).

The most characteristic of these features is the smallness of the disturbed area considering the intensity of the shock at the epicenter. In Great Britain we have no shocks to be compared in strength with those of Fondo Macchia and Ischia. The strongest of British earthquakes are just capable of causing slight damage to buildings (corresponding to the degree 7 of the Mercalli scale), and the average area disturbed by them is 65,900 square miles. Others of intensity 6, only slightly stronger than the Albano

earthquake of 1906, are felt over an average area of 24,500 square miles.

As will be seen from Figure 8, the rapid decline in the intensity of the shock is due to the shallowness of the focus. The curves marked *A* and *B* in Figure 8 represent the intensities of the shock at different distances from the origin, *A* corresponding to a focus $\frac{1}{4}$ of a mile in depth, *B* to one at a depth of 2 miles. The curves are drawn on the assumptions that the intensity of the shock at any point of the surface varies inversely as the square of the distance from the focus, and that the impulses are such that the intensity of the shocks vertically above both foci is the same.

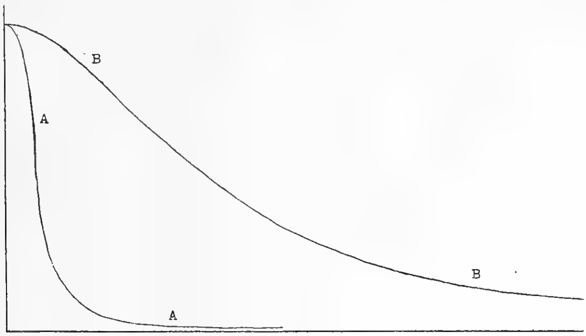

FIG. 8.—Diagram illustrating shallowness of foci in volcanic earthquakes

On these assumptions it follows that the shock that would just be perceptible at a distance of 2 miles from the center for the shallow focus would be perceptible at a distance of 32 miles for the deeper one—that is, the disturbed area of the shock with the deeper focus would be 256 times the disturbed area of the other shock.

It follows, then, that the foci of most volcanic earthquakes are situated at a very slight depth below the surface, and further that the foci vary in depth, the focus of an earthquake like the Nicolosi earthquake of 1901 or the Fondo Macchia earthquake of 1911 being much nearer the surface than that of the Albano earthquake of 1906.

This inference, as to the shallowness of the foci in volcanic earthquakes, is supported by two other lines of evidence. In the

Ischian earthquakes of 1881 and 1883 the depth was estimated, from the inclination of the fissures in walls, to be $\frac{1}{3}$ mile. Again, the duration of the preliminary tremor of an earthquake increases with the distance from the focus, and the practical absence of any such tremor therefore points to the nearness of the focus to the surface.

We may thus conclude: (*a*) that the foci of volcanic earthquakes are situated at a very slight depth below the surface; (*b*) that the foci are usually small and seldom more than 4 or 5 miles in length; (*c*) that the after-shocks, when they occur, originate chiefly within the focus of the principal earthquake; and (*d*), from the discussion of the Etnean and Alban earthquakes, that, while the majority of volcanic earthquakes originate along radial fractures of the mountain, some, and by no means the least important, originate along perimetric fractures.

III. THE ORIGIN OF VOLCANIC EARTHQUAKES

The earthquakes which occur in the neighborhood of active or dormant volcanoes are naturally attributed to the processes which sooner or later tend to result in an eruption. As any sudden displacement of material within the earth's crust may give rise to an earthquake, the processes which have been suggested as possible causes may be grouped under the following heads:

a) The formation of new fractures, or the reopening or extension of old fractures, in the mountain mass, due to the pressure of the column of lava or of gaseous materials in their progress toward the surface.

b) Explosions of any kind, such as those from the sudden generation of steam within the volcano by the access of water to the highly heated rocks below.

c) The sudden injection of fluid rock into fractures or cavities formed in the mass of the volcano.

d) The slipping of the rock surfaces adjoining a fracture due to subterranean movements of the magma.

All of these are possible causes of volcanic earthquakes, and all four processes may be in action during or near the time of an eruption. The question we have now to consider is which cause

must be appealed to as offering the best explanation of the various phenomena of volcanic earthquakes.

For a long time the rending of the mountain mass has been regarded as a probable cause of volcanic earthquakes. G. P. Scrope wrote:

> It may very plausibly be suspected that each of the shocks by which the environs of a volcano are so repeatedly agitated, during, and previous to an eruption, is occasioned by the rending of some part of the solid frame-work of the mountain or its supporting strata, by the action of the force—resulting from the pressure in all directions of the liquid which is in communication with that elevated within the volcanic chimney. The prolongation or widening of a fissure previously formed would have the same jarring, or vibratory effect, as the creation of a new one.[1]

Nearly half a century later[2] Scrope repeated this paragraph with hardly any change except in the initial words, and these were altered to "there can be little doubt that," etc. That the fracturing of the mountain mass would be attended with considerable noise and some perceptible vibration cannot be disputed. Whether it is competent to produce earthquakes so strong as those which visit the southeastern flank of Etna or to give rise to the earthquake phenomena described above is less clear and remains, I think, unproved.[3]

Professor Omori ascribes volcanic earthquakes without explosions to the underground expansive force which produces cracks at the depth of a few kilometers; and those with explosions partly to the upward extension of the cracks, and partly, as regards the slow movements, to the outward pushing of the mountain mass consequent to the explosion.[4] If this be the case, it would seem that the earthquakes due to explosions would usually be of slight

[1] *Considerations on Volcanos* (1825), p. 155.

[2] *Volcanos* (1872), p. 163.

[3] The explanation has received support from well-known seismologists, among others from G. Mercalli (*Vulcani e Fenomeni Vulcanici in Italia* [1883], pp. 354–55), J. D. Dana (*Characteristics of Volcanoes* [1890], p. 22), F. Omori (*Bull. Imp. Earthquake Inv. Com.*, Vol. VI [1912], pp. 132–33), and A. Riccò (*Boll. Soc. Sis. Ital.*, Vol. XVI [1912], p. 32), but not from H. J. Johnston-Lavis (*Monograph of the Earthquakes of Ischia* [1885], p. 92).

[4] *Bull. Imp. Earthquake Inv. Com.*, Vol. VI (1912), pp. 132–33.

intensity, for less than 1 per cent of those recorded on the Asama-yama were sensible to the observers on the mountain side. Here, again, we have no doubt a true cause of volcanic earthquakes, but one of unproved competence to produce the destructive shocks that visit the flanks of Etna and Monte Epomeo.

The sudden injection of lava into fissures and cavities would seem to be more effective and was regarded by the late Dr. H. J. Johnston-Lavis as the cause of the Ischian earthquakes. The theory may be held to account for the extreme localization of the earthquakes, for their repetition time after time in the same region, and for the brevity and sudden onset of the shock.[1]

Within the last few years the view has been gradually gaining ground that many of the stronger volcanic earthquakes are in reality tectonic earthquakes precipitated by volcanic action, that they are due not so much to the actual formation of fractures within the mass of the mountain as to slipping along pre-existing fractures. This view has the merit of connecting the earthquakes under extinct volcanoes such as the Alban Hills, with those, from which they do not in reality differ, under active or dormant volcanoes.

Professor G. Platania has shown that the Linera earthquake of May 8, 1914, was connected with a pre-existing fracture of the volcano, and he urges that subterranean movements of the magma have produced powerful tensions in the eastern region of Etna, by means of which the strata have been fractured, or pre-existing fractures have been enlarged, or the strata have slipped along them, and that these are the causes of the earthquakes.[2]

In some of the mining districts of Great Britain local earth-shakes occur which resemble the Etnean earthquakes so closely in their nature, and probably also in their origin, that it may be useful to give a brief description of one of them.

In Figure 9 are shown the isoseismal lines of an earth-shake which occurred at Pendleton (near Manchester) on November 25, 1905. The intensity of the shock was as high as 7 according to the Rossi-Forel scale (or 6 according to the Mercalli scale). The shock was felt over an area of 144 square miles. As the average

[1] *Monograph of the Earthquakes of Ischia* (1885), pp. 89–95.

[2] *Pubbl. dell'Ist. di Geog. Fis. e Vulcan. della R. Univ. di Catania*, No. 5 (1915), p. 41.

disturbed area of British earthquakes of the same intensity is 24,500 square miles, it is evident that the earthquake originated at a very slight depth. The mean direction of the longer axes of the isoseismal lines is N.37° W. That of the Irwell Valley fault near the epicenter (represented by the broken line on the map) is N.34° W. It would therefore seem probable that the

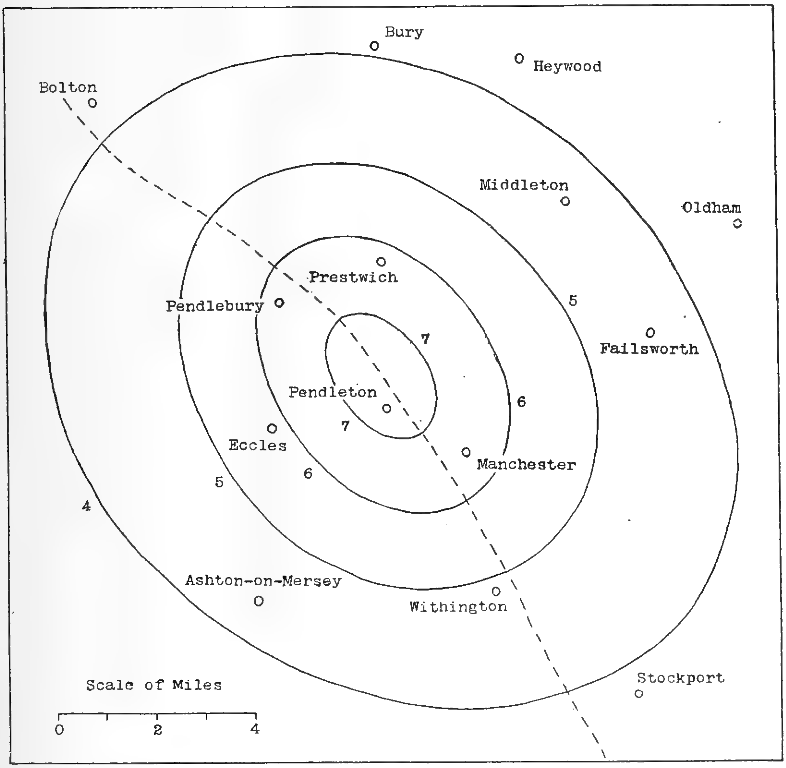

FIG. 9.—Map of Pendleton earth-shake of November 25, 1905

earth-shake was due to a slip along this well-known fault, and at a depth so small that it may well have occurred within the portion of the crust penetrated by the pit-workings at Pendleton. If this be the case, the fault-slip must have been precipitated, not by the natural and gradual growth of the fault, but by the removal of the rock which, it is known, has been effected right up to the

fault, or, possibly, by the pumping of water from the mine.[1] In other words, the Pendleton and other similar earth-shakes are of natural origin in so far as they are due to the growth of faults, but of artificial origin in that the slips are precipitated by mining operations.

An active volcano is coursed by many fractures, the majority of which are radial, but a few perimetric. If the rock mass or masses were in some way to be deprived of support, one mass would slide against the other, and the friction of the grating surfaces would give rise to an earthquake, not to be distinguished, except by its scale, from a true tectonic earthquake. In what way or ways could the rock mass be deprived of its support? Professor Platania suggests by underground movements of the magma,[2] and this might evidently be the case beneath active, and possibly beneath dormant, volcanoes. In all kinds of volcanoes, and especially in extinct volcanoes, another cause may be in operation, though acting much more slowly and giving rise therefore to infrequent shocks like those of Ischia. This is the gradual cooling of lava or heated rock beneath the fractured mass. And it should be noticed that the intensity of the resulting shock would not depend so much on the distance to which the support is withdrawn as on the area of the focus and on the weight of the rock displaced.

This theory, it will be seen, accounts for all the known phenomena of volcanic earthquakes—for their close connection in space and time with many eruptions, the shallowness of their foci and the great intensity of the shock within a limited area, the small size of the foci and the brief duration of the shocks, the occurrence of series of fore-shocks and after-shocks limited in duration and in space chiefly to the original focus, and, lastly, for the occurrence of volcanic earthquakes beneath dormant and extinct, as well as beneath active, volcanoes.

If this is the correct explanation of the cause of the more important volcanic earthquakes, it follows that such earthquakes are of tectonic origin in so far as they are due to the growth of faults, but of volcanic origin in that the slips are precipitated by present or past volcanic operations.

[1] *Geol. Mag.*, Vol. III (1906), pp. 171–76.

[2] *Pubbl. dell'Ist. di Geog. Fis. e Vulcan. della R. Univ. di Catania*, No. 5 (1915), pp. 1–2, 41.

THE STRATIGRAPHIC AND FAUNAL RELATIONSHIPS OF THE MEGANOS GROUP, MIDDLE EOCENE OF CALIFORNIA

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CONTENTS

INTRODUCTION

The problems connected with the divisions of the marine Tertiary of the West Coast are complicated and their solutions difficult. This is due to the original conditions of deposition and to the magnitude of the crustal movements which have affected all the formations of this province, including the Pleistocene. The sediments were, for the most part, deposited in geosynclinal troughs, some of which appear to have been rather local. The deposits are of enormous thicknesses as compared with those of the same period in the Gulf and Atlantic Coast provinces. The aggregate thickness of the Tertiary of the West Coast exceeds 40,000 feet. Clastic materials predominate, and, over wide areas, the beds are either unfossiliferous or the preservation of the fossils poor, due to the leaching out of the original material of the shells. The destructive leaching is more general in the marine Tertiary of California than in that of Oregon and Washington. Crustal movements occurred along the West Coast more or less interruptedly throughout the Tertiary, and the groups of strata representing the major epochs of this time are, as a rule, separated by angular unconformities. The intense folding and faulting which accompanied these movements, together with the paucity and poor preservation of the faunas, have made the problems of correlation difficult, and it is only by detailed mapping, combined with careful paleontological work, that we can hope ultimately to arrive at the final solution. Due in part to these conditions and also to the fact that workers have been few, little has been accomplished toward distinguishing minor faunal divisions in the Tertiary of the West Coast, i.e., such faunal divisions as

might be found in a group of beds belonging to one epoch of deposition.

The foregoing remarks have especial application in connection with the faunas of the marine Eocene of the West Coast, where the larger part of the work of differentiating them remains to be done. Until very recently, one of the largest and most important breaks in the Tertiary of California, an unconformity in the Eocene deposits, was entirely overlooked, and a considerable part of the marine Eocene section of the state, as well as of Oregon and Washington, has been described in a sequence the reverse of the fact.

The fauna of Dr. R. E. Dickerson's Siphonalia sutterensis zone, regarded by him as a part of the Ione formation in the foothills bordering the western front of the Sierra Nevadas, has been regarded until recently as coming from the uppermost Eocene of the West Coast. The beds of this horizon can be shown to belong well down in the Eocene section, and strata of the same age are found in the Coast ranges of California overlain unconformably by beds of Tejon age. The latter formerly were considered to be the older.

In a recent paper the writer has given a summary of his work[1] on the Eocene in the vicinity of Mount Diablo, and has presented evidence to show that there are at least three major stratigraphic and faunal divisions in the Eocene of this region. Previously, only two such divisions had been recognized. The beds of the new division, to which the name Meganos Group was applied, formerly were considered a part of the Tejon.

As will be brought out below, the writer is not the first to advocate a threefold division of the Eocene of the West Coast. The important fact to be remembered in this connection is that, while the fauna of the Meganos had been recognized by previous workers in this field as belonging to a distinct division of the Eocene, the Ione of Arnold and Hannibal and of Waring, the stratigraphic position of this horizon was wrongly determined and, instead of representing the uppermost Eocene of the West Coast, this division comes below the Tejon, and is the middle of the three Eocene divisions as now recognized.

[1] Bruce L. Clark, "Meganos Group, a Newly Recognized Division in the Eocene of California," *Bull. Geol. Soc. America*, Vol. XXIX (1918), pp. 281–96.

HISTORICAL SKETCH

The most detailed work on the Eocene of the West Coast, since the time of Gabb, is that of Dickerson.[1] He recognized only two stratigraphic units as belonging to this period, the Martinez (Lower Eocene) and the Tejon (Upper Eocene). As the result of his studies of the so-called Tejon, he outlined four faunal zones, thought to occupy four successive horizons in that group. These, beginning with the lowest, were the Turbinolia zone, the Rimella simplex zone, the Belanophyllia variabilis zone, and the Siphonalia sutterensis zone. The first three zones were described from the Eocene section in the vicinity of Mount Diablo, and the fourth from the Eocene bordering the foothills of the Sierra Nevadas, the beds of which apparently form part of the Ione formation.

In the paper in which Dickerson first outlined his ideas of the faunal succession of the Tejon we find the statement:[2]

> A study of the relationship between zone 3, Mount Diablo region, and the Siphonalia sutterensis zone and their geographic position suggest that the uppermost strata of the Marysville Buttes and Oroville were deposited by a transgressing sea, and that only in favored places along the western borders of the Sierra have the latest Eocene sediments been preserved from erosion. Lava caps such as that of the older Basalt of South Table Mountain have preserved these youngest Tejon sediments which have heretofore been regarded as Ione.

The first writers to suggest that there are more than two distinct stratigraphic units or groups in the Eocene of the West Coast were Dr. Ralph Arnold and Mr. Harold Hannibal in a joint review of one of Dr. Dickerson's papers. These writers, in their study of the Eocene of Washington and Oregon, recognized three horizons, the Chehalis (at the base), Oliqua, and Arago or Ione (at the top) formations, all higher than the Martinez (Lower Eocene). Their Chehalis horizon they referred to the

[1] R. E. Dickerson, "Note on the Faunal Zones of the Tejon Group," *Univ. Cal. Pub. Bull. Dept. Geol.*, Vol. VIII (1914), No. 2, pp. 17–25; "Stratigraphy and Fauna of the Tejon Eocene of California," *Univ. Cal. Pub. Bull. Dept. Geol.*, Vol. IX (1916), No. 17, pp. 262–534, Pls. 36–46.

[2] R. E. Dickerson, "Note on the Faunal Zones of the Tejon Group," *Univ. Cal. Pub. Bull. Dept. Geol.*, Vol. VIII (1914), No. 2, p. 24.

same period of deposition as the typical Tejon; they did not recognize their Oliqua in California; the Arago formation was correlated with the Ione along the Sierra Nevada front. Arnold and Hannibal agree with Dickerson in placing the marine beds of the Ione as uppermost Eocene in age, but disagree in that they considered this horizon to belong to a distinct epoch of deposition. The following statements are taken from the paper of Arnold and Hannibal[1]:

> The writers have shown that in Oregon and Washington the Eocene may be divided into three faunal divisions, the Chehalis, Oliqua, and Arago or Ione formations. The Chehalis formation is characterized especially by *Venericardia hornii* Gabb, *Meretrix californica* Gabb, and an austral flora, the Oliqua formation by *Pecten* (*Chlamys*) *landesi* or *Venericardia hornii* Gabb and a tropical flora, and the Arago or Ione formation by *Turritella merriami* Dickerson, a form of *V. hornii* with obsolete ribs (var. *aragonia* A. and H.), and a tropical flora.
>
> The Arago or Ione beds represent a horizon younger than any Tejon recognized in the Tejon or Puget Basin. The Arago or Ione beds occurring as they do in basins distinct from those in which the Tejon series is developed, and being formed at a different period, must be treated as a distinct division of the Eocene.

Professor C. E. Weaver in his study[2] of the Eocene sections of the Cowlitz River Valley, Washington, the section studied by Arnold and Hannibal and where they described their Chehalis horizon as being below the Oliqua, disagrees with them as to the sequence. His stratigraphic study of this section apparently shows that the beds of the Oliqua formation are below those of the Chehalis; in other words, Arnold and Hannibal had their section upside down, a condition similar to that which existed in California.

The first published announcement by the writer of his discovery that there are three distinct groups of strata in the Eocene section of Mount Diablo appeared in a paper to which reference has been made.[3] It was not until this paper was in page proof

[1] Ralph Arnold and Harold Hannibal, "Dickerson on the California Eocene," *Science*, new series, Vol. XXXIX (1914), No. 1016, p. 607.

[2] C. E. Weaver, "Eocene of the Lower Cowlitz River Valley of Washington," *Proc. Cal. Acad. Sci.*, 4th ser., Vol. VI (1916), Nos. 1, 2, and 3, pp. 1–17.

[3] Bruce L. Clark, "Meganos Group, a Newly Recognized Division in the Eocene of California," *Geol. Soc. Amer.*, Vol. XXIX (1918), pp. 281–96.

that sufficient evidence was obtained to show that the fauna of the Meganos belongs to the same epoch of deposition as the Eocene marine beds in the vicinity of Marysville Buttes and at Table Mountain near Oroville. These beds generally have been referred to the Ione formation, and considered to represent the uppermost Eocene of the West Coast. A statement to this effect was included in a note at the end of the paper and in the summary of conclusions.

Briefly summarized, the most important results brought out in the paper last mentioned are: that in the Eocene of the region of Mount Diablo there are at least three groups instead of two, as was formerly believed; and that the beds of this newly recognized epoch of deposition come between those of the Martinez (Lower Eocene) and those of the Tejon (Upper Eocene). It was found that the Meganos Group on the north side of Mount Diablo has a maximum thickness of nearly three thousand feet. These beds previous to this time had been generally mapped as Tejon. The unconformity between the Meganos and the Tejon in this area is such a marked one that it may be classed as one of the major breaks in the Tertiary of the Coast ranges. In places there is a difference of from 15° to 20° in strike, and a maximum difference of about 18° in dip between the beds of the Meganos and those of the Tejon above. It was also the conclusion of the writer that there is a marked difference between the fauna found in the beds below and that in the beds above this unconformity, and finally that the beds of the Meganos Group have a fairly wide distribution throughout the Coast ranges of California, in some localities having been mapped as of Martinez, and in others of Tejon, age.

PURPOSE OF PAPER

The purpose of this paper is to sum up the results of more recent studies of other Eocene sections in different parts of the Coast ranges, including the Mount Diablo section, which was described somewhat in detail in the first paper. The most important contribution is the evidence which shows that there are three general divisions in the Eocene of California, namely, the Martinez, Meganos, and Tejon, each of which may be considered as belonging to a distinct epoch. After discussing the various Eocene sections, reasons will be given for correlating the beds referred to the Meganos

Group in these different areas in the Coast ranges with one another and with the marine Ione formation in the Sierra Nevada foothills,

Fig. 1.—Outline map of California. The numbers indicate the locality of the different sections discussed in this paper: (1) Mount Diablo region; (2) region to the north of Coalinga; (3) south end of San Joaquin Valley; (4) Camulus quadrangle; (5) Table Mountain in the vicinity of Oroville.

as mapped and described by Lindgren and Turner, not, however, including the type section of the Ione.

THE MEGANOS GROUP IN THE VICINITY OF MOUNT DIABLO

Mount Diablo is a much faulted anticline of which the Franciscan series forms the core. The Shasta-Chico series (Lower and Upper Cretaceous) is represented in this section by more than ten thousand feet of shales and sandstone, on top of which, on either side of the anticline, are Eocene strata having a maximum thickness of nearly four thousand feet. These beds in turn are overlain by beds referable to the Oligocene, Miocene, and Pliocene. The two Eocene sections, the one on the north side of the Mount Diablo anticline and the other on the south side, will be considered separately as their outcrops are disconnected.

MEGANOS TO THE NORTH OF MOUNT DIABLO

The type section of the Meganos is on the north side of the Mount Diablo anticline. In this paper, the description of that section is largely a repetition of the data presented in the former paper, but adds some details.

The section extends from about one mile to the west of the old coal-mining town of Nortonville, east and a little to the south of the eastern edge of Mount Diablo Quadrangle. These beds, including the Martinez, Meganos, and Tejon groups, dip to the north, the angle of dip varying from 15° to 40°. The greatest width of the outcrops is about two and a half miles.

Stratigraphy and lithology.—The beds of the Meganos Group in this area rest unconformably on those of the Martinez Group. This unconformity, as stated above, was first described by Dickerson.[1] The lower Tejon, as recognized at that time, is the base of the Meganos, as described in this paper. The Meganos beds in this area have a maximum thickness of about three thousand feet. The section may be roughly divided into five lithologic members. Beginning at the base, these will be designated divisions A, B, C, D, and E.[2]

[1] R. E. Dickerson, "The Stratigraphic and Faunal Relations of the Martinez Formation to the Chico and Tejon North of Mount Diablo," *Univ. Cal. Pub. Bull. Dept. Geol.*, Vol. VI (1911), No. 8, pp. 174–76.

[2] The lower part of this section, that is, the Martinez and divisions A, B, and C of the Meganos, are best exposed in the section just to the west and south of the old town of Stewartville. The upper Meganos beds are best exposed on the ridge just to the north of Deer Valley and to the north of that ridge. The best Tejon section is to be found in the vicinity of the old town of Nortonville; also a very good section may be seen in the vicinity of the old town of West Hartley.

Summary of lithology of section.—The outline below is a generalized section of the Eocene groups as found on the north side of Mount Diablo. The Martinez portion of the section is copied from Dickerson's paper.[1]

The conglomerates at the base of the Meganos Group, division A, because of their peculiar character are worthy of mention. In the vicinity of Stewartville great angular slabs of fossiliferous Chico sandstone, some of them five, six, or more feet in length, are associated with well-rounded quartzitic and igneous bowlders. With them are numerous smaller limestone and sandstone pebbles, derived either from the Martinez or the Chico, showing that these beds are a true basal conglomerate. When followed to the east these conglomerates thin out rapidly.

The shales of division C of the Meganos Group are especially noticeable in that they are so different from anything found in the Tejon series on either the north or south side of Mount Diablo. The dark color, the calcareous lenses and nodules, the surface slaking into small fragments, the presence of carbonaceous material, and the layers of coarse sandstone which separate the different shale members all are similar to lithological characters of the

OUTLINE OF EOCENE GROUPS

		Feet
Tejon Group	6. Clay shales with minor amount of sandstone	500
	5. Fine, buff-colored sandstone; in places hard, calcareous layers contain marine fossils	175
	4. Sandy shales; exposures poor; soil very red	175
	3. Light gray to white, angular-grained sandstones, coarse to medium in texture; cross-bedding common, with minor layers of chocolate-colored shales; two important coal layers	75–400
	2. Chocolate-colored shales, ashy in places, with thin lenticular layers of coarse sandstone; coal layer locally known as Black Diamond vein	50
	1. Conglomerate	0–20

Unconformity

[1] R. E. Dickerson, "Fauna of the Martinez Eocene," *Univ. Cal. Pub. Bull. Dept. Geol.*, Vol. VIII (1914), No. 6, p. 71.

Group	Division	Description	Feet
Meganos Group	E	Clay shales and sandstones at top, grades down into fine, massive, poorly indurated sandstone; exposures of the beds of this division are very poor	0–1,500
	D	Sandstone of medium texture; thin-bedded near bottom; more massive at top; yellow brown to gray in color with about 100 feet of cross-bedded sandstone, eolian type. The massive sandstones near top contain lenses of harder calcareous and fossiliferous sandstones	0–300
	C	5. Dark slate-gray shales; bedding planes fairly distinct; light calcareous nodules and lenses	0–230
		4. Sandstone, fine to coarse in texture; in places forms a grit; contains thin clay lenses; in places contains considerable carbonaceous material	110
		3. Dark slate-gray shales, similar to (1) and (5)	90
		2. Sandstone, medium to fairly coarse; weathers on surface a rusty brown; grains chiefly of quartz and mica	50
		1. Dark slate-gray clay shale; bedding planes indistinct; carbonaceous material abundant	75
	B	Coarse to medium fine, quartzitic, gray to gray-brown micaceous sandstones, with some fine conglomeratic layers which are fossiliferous in the eastern part of the area. Sandstone quartzitic at one horizon	700
	A	Heavy conglomerates; changes to sandstone along the strike; bowlders composed of quartzites, chert, limestone, and large angular slabs of sandstone, containing typical Chico (Upper Cretaceous) fossils	0–50
		Unconformity	
Martinez Group		5. Gray-green shale	300
		4. Gray-green glauconitic sandstone; Trochocyathus zitteli beds	50
		3. Fine-grained gray sandstone	200
		2. Shales and sandstones	100
		1. Brown conglomeratic sandstone; Meretrix dalli beds	50
		Total?	
		Unconformity	
Chico			

Knoxville shales (Lower Cretaceous or Upper Jurassic), as seen in certain sections of this general area. These sediments probably are shallow-water deposits, perhaps laid down in estaurine or partially land-locked basins.

Divisions B and D of the foregoing sections contain much biotite. This may be best seen in the sandstones, division D forming the ridge on the north side of Deer Valley. In certain layers, biotite is very abundant, the flakes being fairly large. Grains of feldspar also are present. In fact, the beds may be described as arkosic. The basal sandstones, division B, are also micaceous. The fauna obtained from these arkosic sandstones indicates a subtropical temperature; the lithology, taken in connection with the evidence for warm subtropical waters, possibly points to arid conditions on the land.

In the basal chocolate-colored shales of the Tejon, many impressions of leaves, rushes, and fossil wood are found. These beds will undoubtedly yield a large and well-preserved flora. These leaf shales were apparently laid down in marginal marine swamps. The presence of shells of the genus Corbicula, in a layer of sandstone in the shales, testifies to brackish or fresh-water conditions.

The most important of the coal beds of this region, and one mined throughout most of the area, is found near the top of these lower Tejon shales. In the vicinity of Nortonville this bed is known as the "Black Diamond Vein." It is reported to have a maximum thickness of about four feet. Above this coal seam at Nortonville is a sandy, conglomeratic bed varying from 1 to 3 feet in thickness, which is highly impregnated with limonite. Rush and leaf impressions were found also in this layer. The close association of this bed with the leaf shales and coal, together with the fact that the limonite deposit is limited to a definite layer over a considerable area, suggests a primary rather than a secondary origin.

The coarse, cross-bedded, light-colored sandstones immediately above the shale may well have been deposited under somewhat similar conditions. Two of the important coal-layers, the "Little" and "Clark" veins, mined for many years at Nortonville and Somerville, are found in these sandstones. The coal of the Clark vein, which is about two and one-half to three feet in thickness,

as exposed in the mine at Nortonville, is intercalated between the coarse, white, quartzitic sands without a trace of shale.

Evidence for unconformity between Meganos and Tejon.—The most important evidence for unconformity between the Meganos and the Tejon is the great difference in strike between the beds of the two horizons, seen at numerous localities; this is very noticeable at the coal mine at Stewartville, where the difference approximates 15° (Fig. 1). The basal sandstone of division D is here in contact with the Tejon, the thickness of the sandstone being approximately 150 feet. Followed west of Stewartville, the sandstone disappears and the basal beds of the Tejon rest directly on the upper dark-colored shale (division C), and a little west and south of Nortonville the Tejon rests on the first sandstone member below the top of division C. Southeast of Stewartville the sandstone of division D emerges from beneath the Tejon and forms the ridge north of Deer Valley; the shaly sandstones and shales of division E also appear, and within 3 or 4 miles of Stewartville show their maximum thickness, 1,500 feet. In the canyon south of the Star Mine, not much more than a mile from Stewartville, the upper shales of division E are well developed.

Besides this difference in strike and the rapid emergence of the upper Meganos beds from beneath the Tejon, a marked difference in dip was noted at a number of localities southeast of Stewartville. In general it appears that there is a difference in dip between the two horizons throughout the entire length of the area. At the west end of the area southwest of Nortonville there is a maximum difference in dip of 18° between the upper Meganos beds and those of the lower Tejon. In the vicinity of Stewartville the difference approximates only about five degrees, while in the vicinity of West Hartley the difference is between 15° and 20°.

In the western part of the area under discussion, there are heavy conglomerates at the base of the Tejon which in some places have a thickness approximating 20 feet. Here they rest on the dark shale of division C, and at a number of localities a sharp irregular contact was seen, the bedding planes of the shale being cut off by the conglomerate. It is a noticeable fact, also, that there is considerable carbonaceous material at the contact. In

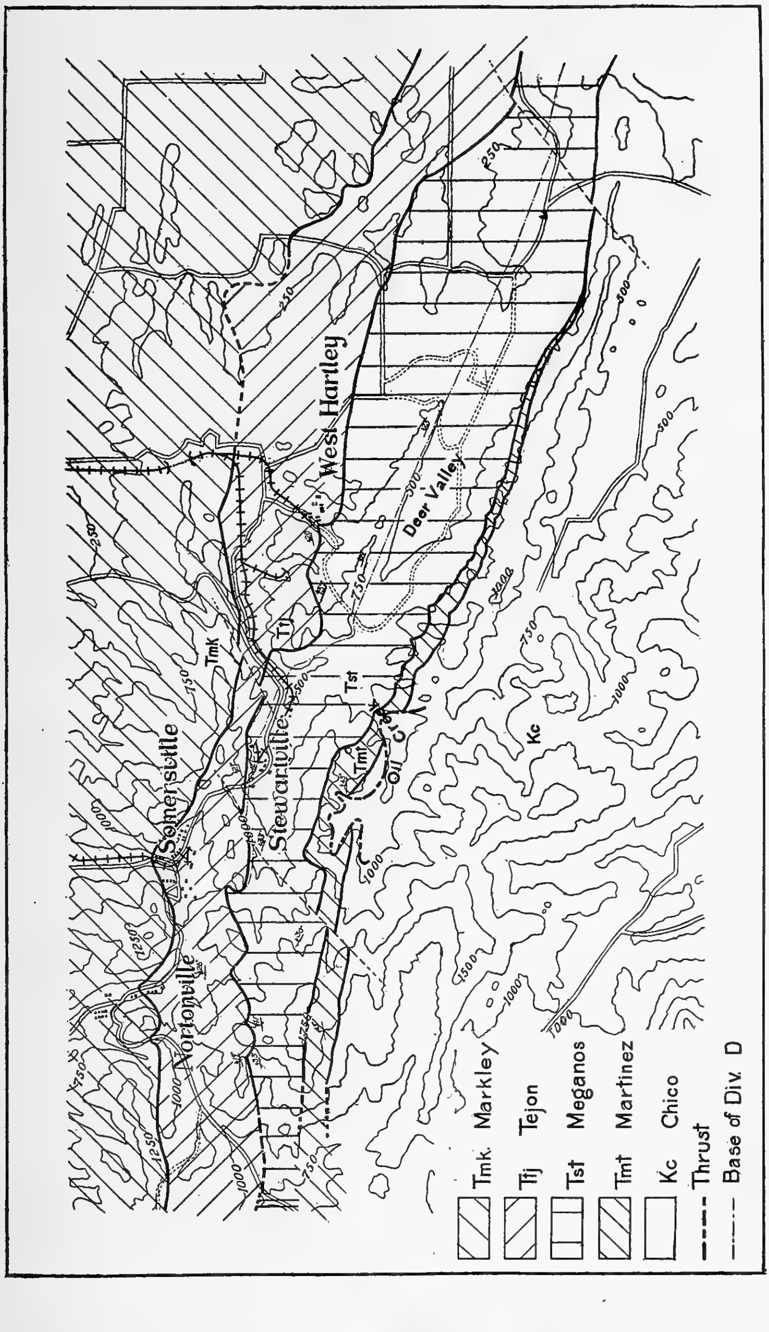


FIG. 2.—Areal map of the Eocene deposits to the north of Mount Diablo

the vicinity of Stewartville and West Hartley the conglomerates disappear and the chocolate shales at the base of the Tejon rest on the shales and shaly sandstones of division E of the Stewartville, making it impossible to find a sharp contact anywhere.

Fauna from Meganos beds.—The following is a preliminary list of the Meganos species obtained from the section described above. The majority of them came from the sandstone division D, a few species from the fine sandstones of division E, and a few from the basal sandstones near the eastern border of the Mount Diablo Quadrangle extending into the Byron Quadrangle to the east:

PELECYPODA

Acila gabbiana Dickerson
Antigona, 2 n. sp.
Arca hornii Gabb, n. var.
Avicula, sp.
Cardium brewerii Gabb, n. subsp.
Cardium marysbillensis Dickerson
Corbula diletata Waring
Corbula, n. sp.
Crassatellites, 2 n. sp.
Dosinia, n. sp.
Diplodonta cretacea Gabb
Glycimeris major Stanton, n. var.
Leda gabbi Conrad
Leda (cf. *alaeformis* Stanton)
Marcia (?) *conradi* Dickerson
Martesia, n. sp.
Macrocallista, n. sp. aff. M. Conradi Gabb
Modiolus ornatus Gabb
Ostrea, sp.
Psammobia, n. sp.
Periploma, n. sp.
Solemya, n. sp.
Solen, n. sp.
Solen, n. sp.
Spisula tejonensis Dickerson
Spisula (cf. *merriami* Dickerson)
Spisula, n. sp.
Tellina longa Gabb
Tellina, sp.
Tellina rêmondii Gabb
Tivela, n. sp.
Venericardia planicosta (cf. var. *merriami* Dickerson)

GASTROPODA

Acmaea, n. sp.
Actaeon, 3 n. sp.
Brachysphingus, 2 n. sp.
Calliostoma, n. sp.
Chrysodomus, 2 sp.
Calyptraea excentrica Gabb
Cancellaria Stantoni Dickerson
Cancellaria, n. sp.
Clavilithes, n. sp.
Conus, sp.
Cylichna, n. sp.
Cypraea, n. sp.
Exilia, sp.
Ficopsis, n. sp.
Fusinus, n. sp.
Galeodea sutterensis Dickerson
Haminea, n. sp.
Natica gesteri Dickerson
Natica hornii Gabb
Natica, n. sp.
Neptunea, n. sp.

Odostomia, n. sp.
Oliva, 2 n. sp.
Phos martini Dickerson
Siphonalia sutterensis Dickerson
Scaphander, n. sp.
Solarium, 2 n. sp.
Terebra, n. sp.
Turris, 5 n. sp.
Turris monolifera Cooper
Turritella, 2 n. sp.
Turritella merriami Dickerson
Whitneya, n. sp.

SCAPHODA

Dentalium (cf. *cooperi* Gabb)
Dentalium, n. sp.

CEPHALOPODA

A Nautiloid, genus indet.

ANTHOZOA

Turbinolia, 2 sp.
Flabellum, n. sp.
Stephanophyllia, n. sp.
Dendrophyllia (?), n. sp.

Comparison of Meganos and Tejon faunas—Up to the present time 68 species have been reported from the Tejon beds on the north side of Mount Diablo. Most of these were listed either by Stanton or Dickerson in the papers already referred to. This upper fauna, referred by Dickerson to his Balanophyllia zone, contains a number of the species which are typical of the type section of the Tejon, such as *Meretrix hornii* Gabb, *Meretrix tejonensis* Dickerson, *Conus remondii* Gabb, *Ficopsis* cf. *cowlitzensis* Weaver, *Turritella uvasana* Conrad, *Turritella uvasana bicarnata* Dickerson.

The fauna of the Meganos as obtained from the type section described above is very different from that of the Tejon. Not more than five of the more than seventy determined species have been found in the Tejon as recognized in this section or known Tejon section. The presence of such described species as *Phos martini* Dickerson, *Siphonalia sutterensis* Dickerson, *Turritella merriami* Dickerson, *Schizaster diabloensis* Kew, together with a fairly large number of undescribed species, is the evidence for correlation of these beds with the Eocene of Marysville Buttes and Oroville, which contain the fauna of Dickerson's Siphonalia sutterensis zone, and with the beds referred to the Meganos on the south of Mount Diablo and the other Meganos sections described in this paper.

MEGANOS TO SOUTH AND SOUTHEAST OF MOUNT DIABLO

SECTION NEAR EAST EDGE OF CONCORD QUADRANGLE

When the first paper on the Meganos was published, comparatively little work had been done on the Eocene section south of the Mount Diablo anticline. A brief description was given in that paper of the beds exposed near the east border of the Concord Quadrangle which joins the Mount Diablo Quadrangle on the west. Here a typical Meganos fauna was found stratigraphically below beds containing a Tejon fauna.

Lithology.—In this section the Meganos beds have an approximate thickness of 2,000 feet. At the base there is between 150 and 200 feet of medium-fine yellow-brown sandstone beginning with about 20 feet of basal conglomerate which contains angular bowlders of fossiliferous Chico (Upper Cretaceous) sandstone, together with angular fragments of shale which is very similar to that found immediately below the contact. The upper 1,800 feet of the Meganos consists principally of dark-colored shales, fine shaly sandstone, and fine sandstone. Some of these shales are almost black and contain considerable lignitic material; they are identical in character with the dark-colored shales in the Meganos north of Mount Diablo.

There is a very marked lithological change between the Tejon beds of this section and those of the Meganos. The Tejon beds consist of 2,000 feet of massive, buff-colored quartzose sandstones which weather into cavernous bluffs on the north side of Pine Canyon a little farther east. At what appears to be the base of the Tejon is a narrow band of fine conglomerate made up of quartz and black and red chert, together with angular fragments of shale similar to the shale member immediately below.

Coal has been found at a number of localities in the basal beds of the Tejon, indicating the same general conditions as those recorded by the sediments on the north side of the mountain. In the sandstones near the base the species *Balanophyllia variabilis*, a coral which is common in the beds above the coal in the Tejon on the north side of Mount Diablo, was found in abundance.

This species is one of the most important markers of Dickerson's Balanophyllia variabilis zone.[1]

EOCENE SECTION SOUTHEAST OF MOUNT DIABLO

The Meganos and Tejon beds of the section just described can be followed continuously eastward, barring some cross-faulting, from the area on the Concord Sheet, to that in the Mount Diablo Quadrangle in which Dickerson made his study. In this section data were obtained for the establishment of three of his faunal zones.

The typical Tejon is represented throughout this area by heavy, massive, buff-colored quartzose sandstones, the outcrops of which form a prominent feature of the landscape. The Meganos beds of the more eastern area are composed for the most part of shale and shaly sandstone, with sandstones at the base, a section which is very similar to that just described, about ten miles to the west.

Detailed mapping has failed to show any marked difference in dip and strike between the Meganos and the Tejon in this southern area, such as occurs to the north of the mountain. At a few localities there is an apparent difference in dip between the beds of the two horizons; this, however, could not be verified with certainty, the division being recognized by a sharp change in lithology, and by faunal evidence.

Faunal zones.—The locality at which Dickerson did most of his work in the "Tejon" of Mount Diablo is southeast of the mountain, in the vicinity of Cave Point and Riggs Canyon. Dickerson divided his (so-called) Tejon into three horizons, the faunas of which were referred to as: (1) the Turbinolia zone; (2) the Rimella simplex zone; and (3) the Balanophyllia variabilis zone.[1]

[1] R. E. Dickerson, "Stratigraphy and Fauna of the Tejon Eocene of California," *Univ. Cal. Pub. Bull. Dept. Geol.*, Vol. IX (1916), No. 17, pp. 373–79.

[2] In the former paper referred to above, the writer stated that in this section thereis a marked difference in strike between the Meganos beds and those of the Tejon, and the difference was taken as one of the evidences of the unconformity between the beds of these two horizons. Later work, however, has shown that this apparent difference in strike is, in part at least, the result of faulting. Also it was stated that to the east of this area the Meganos disappeared due to this unconformity. At that time the writer had not recognized that the so-called Tejon beds to the east, as described by Dickerson, were in part Meganos.

Correlation.—Later work has shown that the faunas of the Turbinolia zone and the Rimella simplex zone belong to the Meganos epoch, while the fauna of the Balanophyllia zone represents typical Tejon.[1] A fairly large number of what are believed to be distinctive markers of the Meganos have been found in the beds referred to the lower two zones just mentioned. A few of the more important of these species which have been found in other Meganos localities and may be considered as markers of that horizon are: *Schizaster diabloensis* Kew, *Turbinolia pusillanima* Nomland, *Venericardia* cf. *merriami* Dickerson, *Trochocyathus imperialis* Nomland, *Ancilla (Oliverata) California* Cooper, *Rimella*, n. sp., *Siphonalia sutterensis* Dickerson, *Turritella merriami* Dickerson, *Turritella andersoni* Dickerson. In the beds representing the other zone, equally good evidence was obtained for correlating them with the typical Tejon.

It is interesting to note at this point that at the time Dickerson wrote his paper "Stratigraphy and Fauna of the Tejon Eocene of California," Arnold, Hannibal, and W. A. Waring correlated the Tejon in the vicinity of Mount Diablo with the Ione as recognized by them, which they recognized as an epoch distinct from and later than the Tejon. The faunas collected by Arnold and Hannibal, and by Waring from the Mount Diablo region appear to have come from the basal beds of the Eocene to the south and southeast of the mountain, Dickerson's lower Tejon recognized by the writer as Meganos. The locality from which the original so-called Ione marine fauna was obtained by these writers, with which they correlated the Mount Diablo fauna, was the south side of Table Mountain. This may be considered the type of Dickerson's Siphonalia sutterensis zone, the fauna of which he thought represented the highest horizon of the West Coast Eocene. This horizon is here placed well down in the Eocene, below the Tejon. Thus Arnold and Hannibal and Waring agree with the writer in their correlation of these lower beds in the Eocene section on the south side of Mount Diablo with the Eocene of Oroville, but they erred in regarding their Ione the uppermost Eocene of the Pacific Coast. They erred with Dickerson in their interpretation

[1] The species listed by Dickerson as Rimella simplex is a new species.

of the sequence, and if they had made sufficient collections, would undoubtedly have recognized the proper sequence.

EOCENE SECTIONS IN THE SOUTHERN PART OF CALIFORNIA IN WHICH THE MEGANOS GROUP IS REPRESENTED

In my first paper on the Meganos Group, reference was made to two Eocene sections in southern California, in which Meganos beds are present. One of these sections is north of Coalinga, on the west side of the San Joaquin Valley; and the other is in the vicinity of Simi Hills, Ventura County. During the summers of 1918 and 1919 several weeks were spent in studying these sections and also the Eocene at the south end of the San Joaquin Valley, where the type section of the Tejon is situated. The results of this work showed conclusively that beds of both Meganos and Tejon age are present in all of these areas, and that there is in each an unconformity separating the strata of these two series. The faunas from the Meganos of these three areas are very similar, containing in common a considerable number of highly ornamented species.[1]

THE SECTION NORTH OF COALINGA

Unconformity.—The unconformity between the Tejon and Meganos groups, in the Eocene section north of Coalinga, has been described by several writers,[2] most of whom considered the beds below the unconformity to be of Martinez age, while the beds above were considered to be Tejon. Dickerson[3] expressed the opinion that the Turritella andersoni beds, those here referred

[1] One of the most common species found in the Meganos of southern California is *Turritella andersoni;* in the past these beds have sometimes been referred to as the Turritella andersoni beds. This species is also found in the Meganos to the southeast of Mount Diablo.

[2] J. A. Taff, "Eocene of the Coalinga-Cantua District, Fresno County, California," *Proc. Pal. Soc. America* (1913), p. 127; E. T. Dumble, "Notes on Tertiary Deposits near Coalinga Oil Field and Their Stratigraphic Relations with the Upper Cretaceous," *Jour. Geol.*, Vol. XX (1912), pp. 28–37; Robert Anderson and Robert Pack, "Geology and Oil Resources of the West Border of the San Joaquin Valley, North of Coalinga, California," *U.S. Geol. Survey, Bull. 603* (1915), p. 66; R. E. Dickerson, "Stratigraphy and Fauna of the Tejon Eocene of California," *Univ. Cal. Pub. Bull. Dept. Geol.*, Vol. IX (1916), No. 17, pp. 382–87.

[3] R. E. Dickerson, *Univ. Cal. Pub. Bull. Dept. Geol.*, Vol. IX (1916), No. 17, pp. 382–87.

to the Meganos Group, were of Tejon age. My conclusion was stated as follows:[1]

After studying the fauna from the Turritella andersoni beds, the same material on which Dickerson based his conclusions, the writer was impressed with the fact that there are so few typical Tejon species in this fauna. He does not agree with a number of the specific determinations that Dickerson made, as given in his list from locality 1817. Evidently Dickerson did not consider the possibility of there being a third group coming in between the Martinez and the Tejon, and that if this were so, one might well expect to find a larger number of species bridging the gap between this intermediate horizon and the Tejon than the gap between the Martinez and the Tejon.

Study of the Eocene series to the north of Coalinga showed conclusively that there is an unconformity in this section, and that the fauna obtained from above this contact is that of the typical Tejon, while the fauna below is referable to the Meganos. It is not the purpose in this paper to describe the lithology of the Eocene of this section except incidentally. Anderson and Pack of the United States Geological Survey have already described the lithology of this section[2] in detail. Accompanying their paper is a geologic map of the area. They referred the beds below the contact to the Martinez, and those above to the Tejon. The writer followed this contact nearly 20 miles, from a point near the old station of Oil City, to the Arroyo Honda near the west border of the Coalinga Quadrangle. Good evidence of an unconformable relationship was found along the entire distance.

As seen between the southern end of Domengine Creek and Cantua Creek (Coalinga Quadrangle), the upper beds of the Meganos consist of a white sandstone, which was mapped by Anderson and Pack as a part of the Tejon.[3] The contact between the Meganos and the Tejon comes in between this sandstone and somewhat similar sandstones of the Tejon. It is, as a rule, marked by a conglomerate, and is irregular at numerous

[1] Bruce L. Clark, "Meganos Group, a Newly Recognized Division of the Eocene of California," *Bull. Geol. Soc. Amer.*, Vol. XXIX (1918), No. 2, p. 294.

[2] Robert Anderson and Robert Pack, "Geology and Oil Resources of the West Border of the San Joaquin Valley, North of Coalinga, California," *U.S. Geol. Survey, Bull. 603* (1915), p. 66.

[3] Anderson and Pack, *op. cit.*, p. 66.

localities. The sandstones below the contact, due to the unconformity, thicken and thin very noticeably along the strike. Also, at a number of localities the lower sandstones show a dip and strike appreciably different from those of the Tejon beds above. While these differences amount at the most to only a few degrees, it is sufficient to cause the lower sandstone layers to be cut off obliquely, and on the cliff sections they are seen to abut against the basal beds of the Tejon (Figs. 3 and 4). Other evidence of this unconformity is the fact that numerous bowlders of sandstone, derived from the Meganos beds below, are found in the conglomerate at or near the base of the Tejon (Figs. 4 and 5).

Fauna.—An invertebrate fauna,[1] listed by Dickerson, was obtained from the beds above the unconformity just noted. It is, apparently, of typical Tejon age, containing a considerable number of highly ornamental molluscan species which have not been found in the Meganos.

The fauna obtained from the beds below the unconformity, the "Turritella andersoni beds," is essentially the same as that of the Meganos in the region of Mount Diablo, the ends in both places containing a fairly large number of highly ornamented species in common. The recognizable described species, which have been obtained from this portion of the section, are indicated in the list on page 000.

Correlation.—Dickerson correlated the Turritella andersoni beds, just mentioned—the Martinez (?) as mapped by Anderson and Pack of the United States Geological Survey—with the lowest Eocene southeast of Mount Diablo (see p. 000), the horizon of his Turbinolia zone. He believed that the beds of this horizon were older than the lowest beds in the type section of the Tejon. The writer agrees with both of these conclusions.

MEGANOS AT THE SOUTH END OF THE SAN JOAQUIN VALLEY

General statement.—An important problem which presented itself in connection with the differentiation of the Meganos from the Tejon, was whether any portion of the type section of the

[1] See list given under University of California locality 672. R. E. Dickerson, "Stratigraphy and Fauna of the Tejon Eocene of California," *Univ. Cal. Pub. Bull. Dept. Geol.*, Vol. IX (1916), No. 17, p. 430.

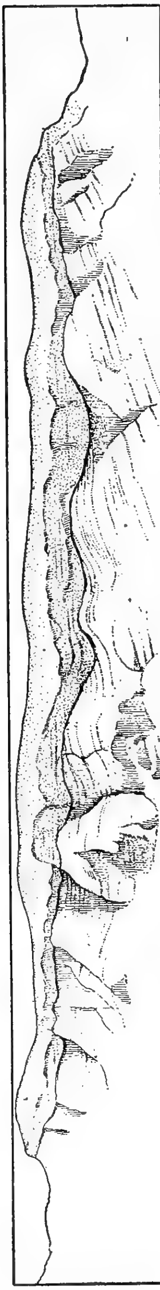


FIG. 3.—Diagrammatic sketch of panorama shown in Fig. 4, illustrating the unconformable contact between the Meganos and the Tejon. The beds, as shown in the view, are dipping into the hill at an angle of close to 30°.

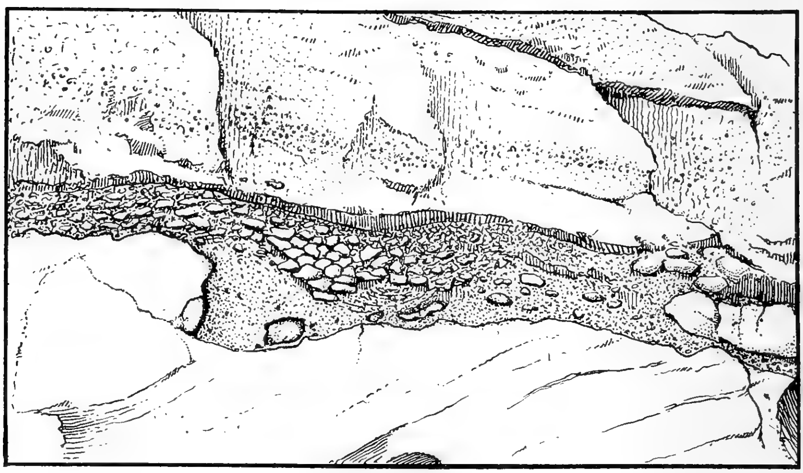
Tejon might be referable to this newly recognized series and, if so, how much of the fauna, previously considered Tejon, belonged in reality to this other division.

The higher mountains immediately south of the San Joaquin Valley, the rocks of which are principally granites and older metamorphics, connect the Sierra Nevadas with the eastern Coast ranges. The hills immediately north of the granitic area and bordering the southern end of the valley are of Tertiary sediments, together with a minor amount of volcanic rock. The oldest unmetamorphosed sediments in this general section are of Eocene age, the outcrops of which may be traced in a narrow belt around the southern end of the valley for a distance of more than thirty miles. These Eocene beds rest on the granites and are overlain unconformably by beds of Oligocene or Lower Miocene age. The Tertiaries in this region have been folded and faulted and the beds as a rule dip at a high angle; in fact, in some localities the beds are overturned to the north, toward the valley, and the sequence is complicated by thrusting. This, however, is not true of the Eocene strata which border the granites. Here, along a narrow east-west belt for a distance of more than twenty miles, is found a normal section. The type section of the Tejon, in Grape Vine Canyon (the Spanish name is Canada de las Uvas) about thirty miles due south of Bakersfield, comes within this belt.

During the summer of 1919, ten days were spent in mapping and studying these Eocene rocks. While much more work remains to be

Fig. 4.—A panoramic view of cliff section composed of Meganos and Tejon rocks, as seen immediately south of divide between Cantua and Salt creeks, Coalinga Quadrangle. The unconformable contact is near the top of the cliff a little below the line of vegetation. Photographed by Mr. Anthony Folger. See diagrammatic sketch of this section in Fig. 3.

a



b

FIG. 5.—(*a*) A close view of the unconformable contact between the Meganos and the Tejon outcrops at a locality about one mile north of Cantua Creek, Coalinga Quadrangle. Photographed by Mr. Anthony Folger. (*b*) Diagrammatic sketch of contact shown in photograph above. The pebbles and bowlders in the conglomerate were derived from the sandstone immediately below.

done before this section is known in detail, enough data were obtained to show conclusively that both Meganos and Tejon are present. The portion of the Eocene outcrops studied extends from San Emigdeo Canyon east to Live Oak Canyon, the latter being the first canyon to the east of Grape Vine Canyon (see map on p. 151).

Faunal evidence for presence of Meganos.—Just to the east of San Emigdeo Canyon fossiliferous beds were found near the base of the section not far above the granite, and from these beds a good Meganos fauna was obtained.

The following is the list of species from this locality:

Cardium, n. sp.	*Amauropsis alveata* Conrad
Cardium cf. *marysvillensis* Dickerson	*Calyptraea* cf. *excentrica*
Glycimeris, sp. (?)	*Natica hannibali* Dickerson
Leda fresnoensis Dickerson	*Rimella*, n. sp.
Meretrix, n. sp.	*Turritella*, n. sp.
Psammobia, n. sp.	*Scaphander*, n. sp.
Tellina, n. sp.	*Seraps erratica* (Cooper)
Venericardia, n. sp.	

This fauna contains several of the distinctive forms of the Meganos, such as *Leda fresnoensis* Dickerson, *Venericardia* n. sp., *Rimella* n. sp., *Natica hannibali* Dickerson, *Turritella* n. sp. Probably the most distinctive species in this fauna is the *Rimella* n. sp., which is very common in beds of the Meganos from Mount Diablo to southern California.

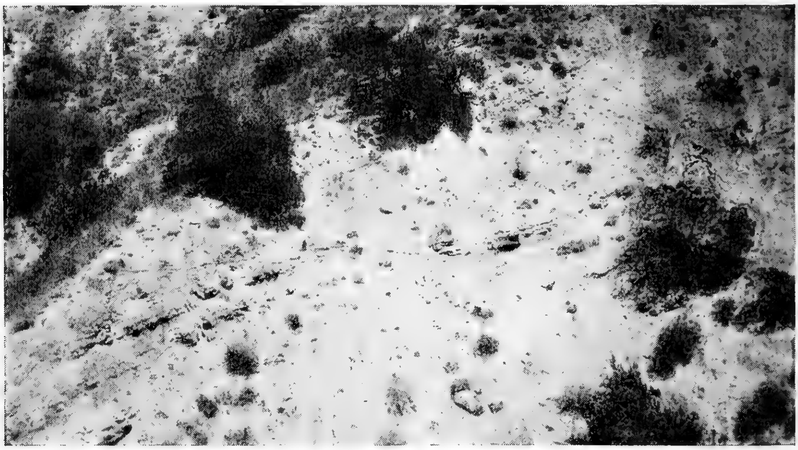
Unconformity.—These fossiliferous Meganos beds were found to rest unconformably below others containing a typical Tejon fauna, the latter connecting directly with the outcrops of the typical Tejon of Grape Vine Canyon.

One of the localities where the unconformable contact between the Meganos and the Tejon may be seen distinctly is about one-eighth of a mile back of the old Douglas ranch-house in the main canyon of San Emigdeo Creek.[1] Here the contact is beautifully exposed on the side of the canyon. There is a difference in dip between the two series of as much as 10° A basal conglomerate containing fossiliferous bowlders derived from the beds below was

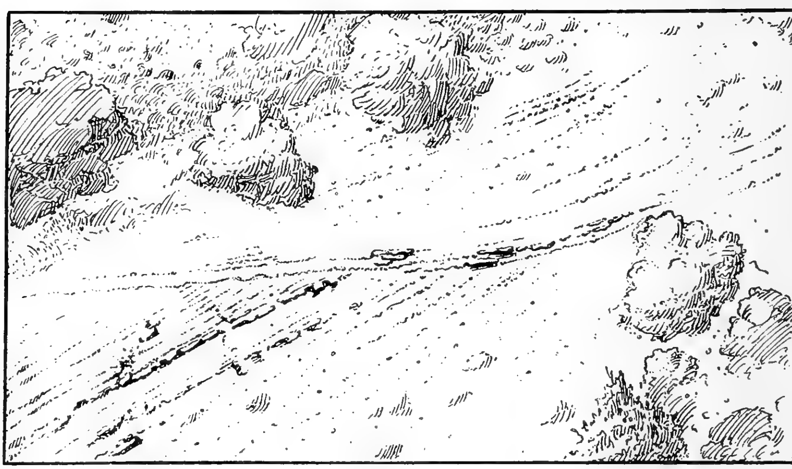
[1] Near south edge of NW. $\frac{1}{4}$, Sec. 5, T. 9 N., R. 21 W.

found along the contact (Fig. 6). The unconformity is also shown on the map (Fig. 7, p. 151).

At the locality just mentioned the Meganos beds have a thickness of less than 15 feet, and not more than 300 feet to the west



a



b

FIG. 6.—(*a*) A close view of the unconformable contact between the Meganos and Tejon outcrops as seen on the east side of San Emigdeo Canyon. (*b*) Diagrammatic sketch of contact shown in photograph above.

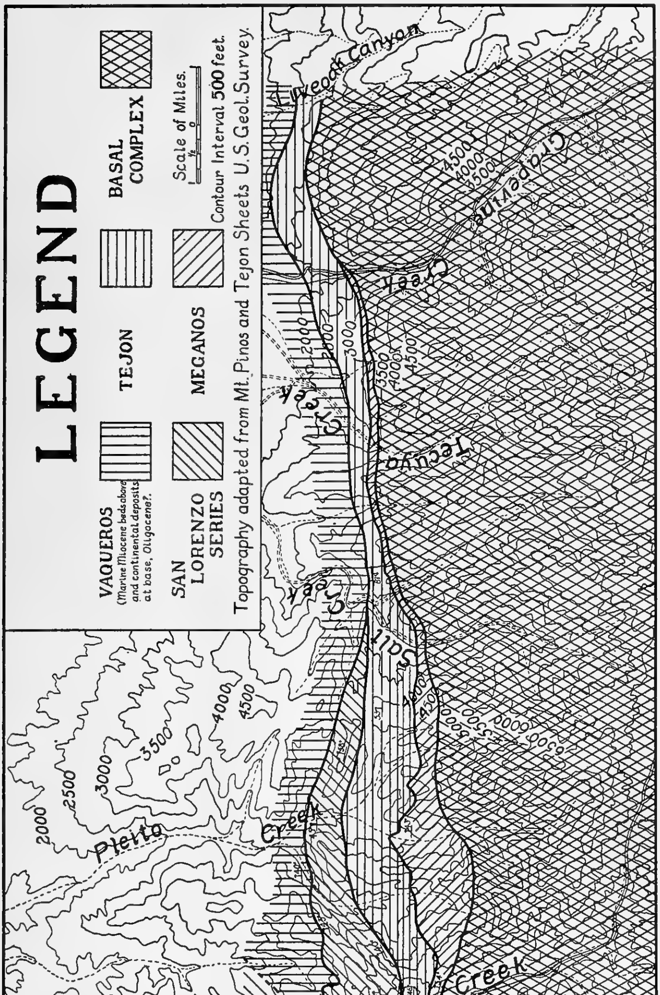


FIG. 7.—Areas map, showing distribution of Eocene and Oligocene outcrops at the south end of the San Joaquin Valley

the basal conglomerate of the Tejon rests on the granite. Just back of the ranch-house, a little farther west, a remnant of the basal Meganos sandstone outcrops below the basal conglomerate. Here the unconformity is evident. Traced east of this locality, the Meganos beds are found to thicken rapidly, reaching their maximum thickness near the head of Pleito Canyon about three miles to the east, where the Meganos beds have an estimated thickness of more than one thousand feet. As shown on the map, these beds thin out rapidly farther east, and in the canyon of Salt Creek, only a little more than four miles distant, their thickness probably is not more than one hundred feet. The conglomerate of the basal Tejon was traced to Tecuya Creek in the next large canyon east of Salt Creek. In Grape Vine Canyon the basal conglomerate of the Tejon was found to be separated from the granite by about twenty-five feet of unfossiliferous, coarse arkosic sandstone, together with a few feet of dark shales.

Thus the beveled Meganos is transgressed by the Tejon from west to east in the vicinity of Grape Vine Canyon, only a very small part of the Meganos being left, and in the next canyon to the east of Live Oak Canyon the Meganos beds fail to appear.

Lithology.—The Meganos outcrops are best exposed between Pleito and San Emigdeo canyons. The basal beds consist of several hundred feet of fairly indurated, coarse, reddish-gray arkosic sandstone. The upper part of the section is composed principally of sandy shales and platy shaly sandstone.

The Tejon beds of this region have a thickness of a little more than two thousand feet. The thickness in the vicinity of Grape Vine Canyon was estimated to be about twenty-four hundred feet. In the vicinity of this canyon the beds consist principally of medium-fine buff-colored sandstone, with lenticular harder, calcareous fossiliferous layers. This section, as already described by Anderson and Dickerson, is very uniform in lithology. To the west these beds become finer, and in the vicinity of San Emigdeo Canyon the larger part of the section might be described as a mudstone. In places lenses of conglomerate are found with the finer sediments, and at one horizon not very far from the base is a heavy layer of conglomerate that can be traced

for a considerable distance. It would appear that the Tejon beds in this last general locality may be delta deposits rather than typical marine deposits, such as those to the east in the vicinity of Grape Vine Canyon. This is borne out by the paucity of the fauna as well as by the lithology.

Fauna of the type Tejon.—The faunas obtained from different horizons in the type section of the Tejon, as found in Grape Vine Canyon, were studied by Dickerson. The invertebrate species were listed and a number of new species described by him.[1] Dickerson's conclusion, with which the writer agrees, was that the fauna obtained from the various horizons in the type Tejon, taken as a whole, is a unit. It has already been pointed out that Dickerson believed that these beds were somewhat younger than the Turritella andersoni beds at Coalinga or his lower Tejon from the south side of Mount Diablo, which beds of both localities are referred by the writer to the Meganos. In discussing this fauna, he says:[2]

> Beds about three hundred feet above the base (University of California locality 458) yielded an excellent fauna. This fauna, however, does not differ essentially from that of the beds higher in the section. The faunas from several other localities which are listed below do not differ materially from one another, but appear to represent one phase only. This faunal unity is in consonance with the sedimentary record as Anderson described it. . . . .
>
> The writer is in complete agreement with Anderson's view as expressed here in relation to the type Tejon. However, beds both higher and lower than the Eocene of Canada de las Uvas occur in other parts of the state, notably in the vicinity of Mount Diablo, along Cantua Creek, Coalinga Quadrangle, and at the Marysville Buttes.

As quoted in the paragraph above, Dickerson recognized that the fauna of the type Tejon was higher than that from the Lower Eocene beds on the south side of Mount Diablo, and higher than his so-called lower Tejon at Coalinga, the Turritella andersoni beds, which latter beds are here referred to the Meganos Group. He correlated the fauna of the type Tejon with that of his Rimella

[1] R. E. Dickerson, "Fauna of the Type Tejon; Its Relation to the Cowlitz Phase of the Tejon Group of Washington," *Proc. Cal. Acad. Sci.*, Vol. V (1915), No. 3, pp. 33–98.

[2] R. E. Dickerson, *op. cit.*, p. 40.

simplex zone. With this correlation I do not agree. As stated in the discussion on p. 000, the species *Rimella simplex* has not been found in the vicinity of Mount Diablo. The specimens from the south side of Mount Diablo, determined as such by Dickerson, belong to a new species which appears to be characteristic of the Meganos horizon. The so-called Rimella simplex beds of Mount Diablo come within the Meganos part of the section, and contain the typical species of that horizon.

EOCENE OF THE CAMULOS QUADRANGLE,[1] VENTURA COUNTY

General.—The fourth Eocene section studied during the summer of 1918 is that of the Camulos Quadrangle of Ventura County, California. The Eocene outcrops are found on both sides of the Simi Valley, the best and most complete section being in the hills on the south side of the valley, the strike of the beds almost paralleling the valley in an east-and-west direction. The late W. A. Waring described and mapped the geology of this area.[2] He recognized two Eocene divisions in this section, the Martinez and the Tejon, stating that apparently the Martinez (Lower Eocene) graded up into the Tejon. The fauna figured and described by him in his paper as Tejon is that of the Meganos. However, the Tejon also is represented in this section resting unconformably upon the Meganos.

Lithology.—This general area is being mapped and described by Dr. William S. W. Kew of the United States Geological Survey. According to him, the maximum thickness of the beds here referred to the Meganos is about three thousand five hundred feet. They consist principally of bluish-gray shales and shaly sandstones. Massive conglomerates are found near but not at the base. No sharp line of division between the Martinez and the Meganos has been found in this section. This, very possibly, is due to the lack of sufficient detailed work. The Tejon here consists of a series of about one thousand five hundred feet of coarse sand-

[1] The eastern half of Camulos Quadrangle comprises the Santa Susana and Calabasas quadrangles.

[2] W. A. Waring, "Stratigraphic and Faunal Relations of the Martinez to the Chico and Tejon of Southern California," *Proc. Cal. Acad. Sci.*, 4th ser., Vol. VII (1917), No. 4, pp. 41–124, Pls. 7–16.

stones, cross-bedded sandstones, and conglomerates. Above this is a great thickness of land-laid beds which are generally correlated with the Sespe formation.

Unconformity.—The contact between the Tejon and Meganos of this section is marked by conglomerates and conglomeratic sandstones. At a number of localities true basal conglomerate was found. The unconformity between the beds of these two horizons is also brought out by the mapping. On the south side of the Simi Valley near its east end the Meganos beds have a thickness of about one thousand five hundred feet; traced westward they thin out rapidly and near the west end of the valley disappear, due to overlap of the Tejon beds. This disappearance of the Meganos beds takes place in a very short distance, there being an appreciable difference in strike between the beds of the two horizons, which could only have been the result of crustal movements.

Fauna.—The following is a list of species obtained from the basal beds of the Tejon of this section, University of California locality 3311:

Cardium brewerii Gabb
Corbicula, n. sp.?
Glycimeris sagitata Gabb
Marcia? n. sp.
Tellina, sp.
Amauropsis alveata Conrad
Crepidula pilium (Gabb)
Ficopsis remondii Gabb
Natica hornii Gabb
Pseudoperissolax blakei (Conrad)
Turritella uvasana Conrad
Turris (*Surculites*) *sinuata* Gabb
Turris (*Surcula*) *io* Gabb
Whitneya ficus Gabb

Though this fauna is a small one, the writer feels confident in his correlation of these beds with those of the typical Tejon, because: (1) of the presence of an angular unconformity between the beds containing these species and those containing a typical Meganos fauna; (2) because it is believed that a number of the species listed above are characteristic of the Tejon. All are very common in the fauna obtained from the type section of the Tejon, and only four of the species have been found in beds of Meganos age: *Amauropsis alveata* Conrad, *Ficopsis remondii* Gabb, *Natica hornii* Gabb, and *Pseudoperissolax blakei* (Conrad).

The fauna obtained from the Meganos of this general section is one of the best preserved and largest from any known section

belonging to that epoch of deposition. A very large percentage of the species are common to the Meganos of the Coalinga section, as well as to that of the Mount Diablo region.[1]

Correlation.—It was from these Eocene beds in Ventura County that convincing evidence was first obtained that the Meganos belongs to the same horizon as that of the Eocene of Marysville Buttes and Table Mountain near Oroville, California, the beds of which localities contain the fauna of the Siphonalia sutterensis zone. The large number of highly ornamented species common to the Meganos of the Ventura County region and to the Eocene of these other localities seems to show conclusively that we are dealing with beds that are nearly, if not exactly, contemporaneous.

One of the localities, from which the writer has obtained the best-preserved Meganos fauna in the Ventura County area, is along Aliso Canyon about four miles northeast of the east end of Simi Valley. Here were found a number of the species which have been regarded as characteristic of the Siphonalia sutterensis zone.[2] The following quotation is taken from the published abstract of one of Dickerson's papers in which he refers to this section:

> A year ago Mr. Reginald Stoner discovered a locality in the Santa Susana Mountains, on Aliso Canyon of Devil Creek, just beneath the Miocene strata. The fossils from this locality represent a lower phase of the Siphonalia sutterensis zone and the fauna is essentially the same as the Siphonalia sutterensis zone of the Roseburg Quadrangle, on Little River near the confluence with the Umpqua.
>
> In the Simi Hills, a few miles away from the locality discovered by Mr. Stoner, the Rimella simplex zone of the middle Tejon stage occurs; the general absence of this zone through most of the Coast Range region is probably due to extensive erosion during the interval between upper Eocene and Oligocene time.

Dickerson, at the time the above-mentioned paper was written, supposed that these beds containing the fauna which he recognized

[1] For the list of the described species from the Meganos of the area under discussion the reader is referred to the list on pages 158–59.

[2] R. E. Dickerson, "Occurrence of the Siphonalia Sutterensis Zone, the Uppermost Tejon Horizon in the Outer Coast Ranges of California," *Bull. Geol. Soc. America*, Vol. XXIX (1917), p. 163.

as that of his Siphonalia sutterensis zone were at the top of the Eocene in this general section, and used this as corroborative evidence in support of his belief that this zone belongs to the uppermost Eocene horizon known on the West Coast. At that particular locality these beds are in unconformable contact with beds of Lower Miocene age. Further stratigraphic work by Kew, however, has shown that to the west other beds come in between these Eocene beds of Aliso Canyon and the Lower Miocene, and not more than four miles from that locality nearly 3,500 feet of other strata are found between. These include beds of true Tejon age, together with a considerable thickness of land-laid beds which have generally been called the Sespe formation. This is the section already referred to, in which there is a marked unconformity between the Meganos and the Tejon. Thus mapping and faunal work in this region show conclusively that the Eocene beds of Aliso Canyon, correctly considered by Dickerson as representing his Siphonalia sutterensis zone, lie unconformably below beds which contain a typical Tejon fauna.

FAUNA OF THE MEGANOS OF CALIFORNIA

The following is a list of the described species from the different Eocene localities in California which are now believed to be of Meganos age. The list is as complete as can be made at the present time. The writer takes entire responsibility of the specific determinations. The fauna, as listed here, represents only a comparatively small part of the known fauna, since a very large proportion of the known Meganos species has not yet been described. When this fauna is more thoroughly worked up, the evidence for the correlation of the different sections here described will appear more conclusive.

The general localities from which the species listed on pages 158–59 have been obtained are indicated in the columns on the side, as follows: M., Marysville; T.M., Table Mountain; Mt.D., Mount Diablo; Coal., Coalinga; Ca., Camulos and Calabasas Quadrangles; T.T., Type Tejon.

TABLE I

	M.	T.M.	Mt.D.	Coal.	Ca.	T.T.
Anthozoa:						
Flabellum (?) merriami Nomland				×		
Stephanophyllia californica Nomland	×		×			
Trochocyathus imperialis Nomland			×	×		
Trochocyathus perrini Dickerson	×		×	×		
Thamnasteria sinuata Nomland			×			
Turbinolia dickersoni Nomland			×	×		
Turbinolia pusillanima Nomland			×	×		
Echinodermata:						
Schizaster diabloensis Kew	×		×			
Pelecypoda:						
Acila gabbiana Dickerson	×	×	×	×		× ?
Arca clarki Dickerson			×			
Arca hornii Gabb, n. var.						
Cardium brewerii Gabb, n. subsp	×		×	×	×	
Cardium marysvillensis Dickerson	×	×	×	×	×	
Corbula diletata Waring						
Crassatellites lillisi Dickerson				×		
Cucullaea morani Waring					×	
Diplodonta cretacea (Gabb)						
Glycimeris fresnoensis Dickerson				×		
Glycimeris marysvillensis Dickerson	×					
Glycimeris major Stanton, n. var.			×			
Isocardium tejonensis Waring	×		×	×	×	
Leda fresnoensis Dickerson				×		
Leda gabbi Conrad			×	×		×
Marcia (?) conradi Dickerson				×		
Modiolus ornatus Gabb	×	×	×	×	×	×
Nucula cooperi Dickerson	×					
Phacordes gyrata Gabb						
Spisula tejonensis Packard	×	×	×		×	
Spisula merriami ?			×			
Tellina sutterensis Dickerson	×					
Tellina longa Gabb						
Tellina rêmondii Gabb						
Tivela weaveri Dickerson					×	
Venericardia planicosta merriami Dickerson	× ?	× ?	× ?			
Gastropoda:						
Acmaea ruckmani Dickerson		×				
Amauropsis alveata (Conrad)	×		×	×	×	×
Ancilla (Oliverata) californica Cooper	×	×	×		×	
Bittium featherensis Dickerson		×				
Bittium longissimum Dickerson	×					
Calliostoma arnoldi Dickerson	×					
Calyptrae excentrica (Gabb)	×	×	×	×	×	×
Cancellaria irelaniana (Cooper)	×	×				
Cancellaria stantoni Dickerson	×	×	×	×	×	×
Caricella stormsiana Dickerson	×					
Cerithiopsis orovillensis Dickerson		×				
Chrysodomus ? martini (Dickerson) = *Phos martini* Dickerson	×	×	×			
Clavilithes tabulata Dickerson	×				×	
Cordiera gracillima Cooper	×	×				
Exilia perkinsiana Cooper	×		×			

TABLE I—*Continued*

	M.	T.M.	Mt.D.	Coal.	Ca.	T.T.
Gostropoda (*continued*):						
Ficopsis rêmondii Gabb, n. var.	×	×	×	×	×	×
Fusinus lineatus Dickerson	×	....	....	....	....	....
Fusinus merriami Dickerson	×	....	....	....	....	....
Galeodea sutterensis Dickerson	×	×	×	....	×	....
Lyria andersoni Waring	....	....	....	×	×	....
Metula harrisi Dickerson	....	....	....	×	....	....
Mitra simplicissima Cooper	×	....	....	....	×	×
Moiopophorus striatus Gabb	....	....	×	....	....	×
Monodonta wattsi Dickerson	×	×	....	....	....	....
Murex nashi Dickerson	×	....	....	....	....	....
Natica gesteri Dickerson	....	....	×	×	....	....
Natica hornii Gabb	....	....	....	....	....	....
Natica subobesa (Cooper)	....	....	×	....	....	....
Natica hannibali Dickerson	....	....	....	×	×	....
Natica nuciformis Gabb	....	....	×	....	....	×
Nyctilochus thunani Dickerson	×	....	....	....	....	....
Olivella marysvillensis (Dickerson)	×	....	....	....	....	....
Pseudoliva dilleri Dickerson	....	....	....	....	×	....
Pseudoperissolax blakei (Conrad), n. subsp.	×	....	×	×	×	×
Seraphs erratica (Cooper)	×	....	....	....	×	....
Siphonalia sutterensis Dickerson	×	×	×	....	....	....
Solarium Weaveri (Dickerson)	×	....	....	....	....	....
Solarium ulreyana Dickerson	×	....	....	....	....	....
Spiroglyphus (?) *tejonensis* Arnold	....	....	×	×	....	....
Strepsidura howardi Dickerson	....	....	×	....	....	....
Terebra wattsiana Cooper	×	....	....	....	....	....
Turris (*Pleurotoma*) *cooperi* Dickerson	×	....	....	....	....	....
Turris (*Pleurotoma*) *monolifera* Cooper	×	×	×	×	×	....
Turris (*Pleurotoma*) *ulreyana* Cooper	×	×	....	....	....	....
Turris (*Surcula*) *clarki* Dickerson	×	×	....	....	×	....
Turris (*Surcula*) *crenatospira* Cooper	×	....	....	×	×	....
Turris (*Surcula*) *davidsiana* (Cooper)	×	....	....	....	....	....
Turris (*Surcula*) *holwayi* Dickerson	×	....	....	....	....	....
Turris fresnoensis Arnold	....	....	....	×	....	....
Turris guibersoni Arnold	....	....	....	×	....	....
Turris inconstans Cooper	×	×	....	....	....	....
Turris suturalis (Cooper)	×	....	....	....	....	....
Turritella andersoni Dickerson	....	....	×	×	×	....
Turritella merriami Dickerson	×	×	×	....	....	....
Voluta lawsoni Dickerson	×	....	....	....	....	....

CORRELATION

CORRELATION OF MEGANOS SECTIONS IN COAST RANGES

The correlation of that portion of the different Eocene sections of the Coast ranges which has been referred to as the Meganos is based on both stratigraphic and faunal evidence. The stratigraphic evidence in itself, without the faunal, would

not be sufficient as it is impossible to trace the beds by mapping from any one of these general localities to another.

The stratigraphic evidence shows that in all the localities which have been examined, the Mount Diablo, Coalinga, and the Simi Hills regions, an unconformity exists between beds containing a typical Tejon fauna and others containing a fauna which is very different from that of the typical Upper Eocene, and also very different from that of the Martinez (Lower Eocene). As has been pointed out these general unconformities are not the result of local crustal movements, and surely cannot be classed as being "at most secondary order, i.e., such as might separate two formations within a group."[1] The beds below the upper unconformity are not Martinez in age, as shown by the fact that in the Mount Diablo region the Meganos beds rest unconformably on the Martinez. It is not possible at this time to present all the faunal evidence for correlating the different sections of the Meganos of the Coast ranges, as a large percentage of the species from this horizon are new and have not been described. The following discussion is based on described species only.

The best faunal evidence for correlating the Meganos of the Mount Diablo region with that of the Coalinga region is that presented by the corals. Three described species of corals are common to these two general sections; these are *Turbinolia pusillanima* Nomland, *Turbinolia dickersoni* Nomland, and *Trochocyathus imperialis* Nomland.[2] It has already been pointed out that Dickerson correlated the beds in the Coalinga region, which are here referred to the Meganos epoch of deposition, with those of his Turbinolia zone as recognized in the Eocene section

[1] R. E. Dickerson, "Stratigraphy and Fauna of the Tejon Eocene of California," *Univ. Cal. Pub. Bull. Dept. Geol.*, Vol. IX (1916), No. 17, p. 429.

[2] J. O. Nomland, "Corals from the Cretaceous and Tertiary of California and Oregon," *Univ. Cal. Pub. Bull. Dept. Geol.*, Vol. IX (1916), No. 5, pp. 59–76, Pls. 3–6.

Dr. Nomland listed *Turbinolia dickersoni* as being present in the Tejon of the Coalinga region; the type, however, came from the Meganos of this same section. Later examinations by Nomland of the specimens from the Tejon of this region, determined by him as *T. dickersoni*, show that this determination was a wrong one and that the form from this horizon is apparently a new species.—R. E. Dickerson, *Univ. Cal. Pub. Bull. Dept. Geol.*, Vol. IX (1916), No. 17, pp. 427–28.

southeast of Mount Diablo. This correlation apparently was based primarily on the corals. Dickerson believed that the beds of his Turbinolia zone represented a horizon lower than that of the type section of the Tejon. Thus, as regards the stratigraphic position of his Turbinolia zone with the true Tejon, he and the writer are in agreement. Besides the corals, there is a considerable number of highly ornamental molluscan species common to the Meganos of the Mount Diablo and Coalinga regions. A very large proportion of these are new species belonging to such genera as *Meretrix*, *Rimella*, *Turris*, *Ficopsis*, *Galeodea*, *Turritella*, etc. One of the described gastropod species, found in all three of the Meganos sections under discussion, is *Turritella andersoni*. This species appears to be a marker of the Meganos horizon.

The evidence for the correlation of the Meganos of the Coalinga region with that of the Camulos Quadrangle in Ventura County is even more conclusive. The faunas of the Meganos of these two localities have a very large proportion of their species in common, while in the more southern locality, Camulos Quadrangle, a number of species are found which are common to the Meganos of the Mount Diablo region, but which have not been found in the Coalinga region; among these are *Ancilla* (*Oliverata*) *californica* Cooper, *Galeodea sutterensis* Dickerson, and *Turritella merriami* Dickerson, none of which, as far as the writer is aware, have been found in the beds of typical Tejon.

CORRELATION OF THE MEGANOS WITH THE SIPHONALIA SUTTERENSIS ZONE OF THE IONE FORMATION

As already stated, it is my conclusion that the Meganos Group, originally described from the vicinity of Mount Diablo, belongs to the same general epoch of deposition as the beds of the Siphonalia sutterensis zone described by Dickerson[1] and which he considered the uppermost Eocene of the West Coast, and a part of the Ione formation described by Turner from the Jackson Quadrangle.

[1] R. E. Dickerson, "Note on the Faunal Zones of the Tejon Group," *Univ. Cal. Pub. Bull. Dept. Geol.*, Vol. VIII (1914), No. 2, pp. 17–25; "Stratigraphy and Fauna of the Tejon Eocene of California," *Univ. Cal. Pub. Bull. Dept. Geol.*, Vol. IX (1916), No. 17, pp. 363–524, Pls. 36–46.

What may be considered as the type section of the Siphonalia sutterensis zone is found at Table Mountain near Oroville, California. Dickerson[1] also referred the Eocene of the Marysville Buttes to this horizon, and later the fauna of the Umpqua formation of the Roseburg Quadrangle was included in the same horizon.[2] He recognized the distinctiveness of the Siphonalia sutterensis fauna from that of the type Tejon. The absence of many of the highly ornamented molluscan species, so common in the typical Tejon, and the presence of an equally large number of species in the former fauna not found in the latter, appear to be the principal reasons for his belief that the two faunas were not contemporaneous. Dickerson attempted to establish the stratigraphic sequence of his upper faunal zone in relation to that of the typical Tejon indirectly, not having the two faunas in the same section. His idea that the Siphonalia sutterensis fauna is younger than that of the typical Tejon appears to have been founded principally upon what he considered evidence for different stages of evolution of certain pelcypods, such as *Venericardia planicosta merriami* Dickerson and *Cardium marysvillensis* Dickerson. He believed that the variety *merriami* was derived from the variety *hornii*. Later stratigraphic work has shown that these species occur in a sequence the reverse of that which Dickerson originally supposed, the *Venericardia planicosta merriami* coming in beds older than those containing the variety *hornii*. The same is true of the other species, which were derived from typical Tejon species.

Another line of evidence which was presented as a basis for believing that the fauna of the Siphonalia sutterensis zone is closely related to that of the typical Tejon and therefore should be classed as Tejon, is the presence of a large percentage of species in the former fauna which, according to his determination, are also present in the typical Tejon. The writer has had access to all the collections which Dickerson had when he came to the foregoing conclusions. Their study has shown that there is a

[1] *Op. cit.*, pp. 403–6.

[2] R. E. Dickerson, "The Fauna of the Siphonalia Sutterensis Zone in the Roseburg Quadrangle, Oregon," *Proc. Cal. Acad. Sci.*, Vol. IV (1914), pp. 113–28, Pls. 11–12.

much smaller percentage of species common to the Siphonalia sutterensis zone and the typical Tejon than was supposed. As might be expected, there are a few species common to both faunas, but taken as a whole they are distinct. This will be still more evident when the entire fauna of the Meganos is described.

While the number of species common to the Eocene of the Marysville Buttes and Table Mountain, and to any one section of the Meganos of the Coast ranges mentioned in this paper (species not found in the typical Tejon), is not very large, yet they are forms such as would not be expected to have a very long range. A number of species from the Siphonalia sutterensis zone have been found in the Meganos of the Ventura County section, which have not been found in the Meganos of the Mount Diablo region, and vice versa. Taken as a whole, as indicated in the list, pages 158–59, there is a fairly large number of distinctive species common to the general Meganos of the Coast ranges and the beds of the Siphonalia sutterensis zone found in the Marysville Buttes and in the vicinity of Oroville. Among these are several corals together with a fairly large number of highly ornamented gastropoda and pelecypoda, the type species which are generally regarded as good horizon markers.

GENERAL CORRELATION

The consensus of opinion of those who are familiar with the Tejon fauna of California has been that it represents about the same stage of deposition as the Claiborne of the Gulf province, which in turn is correlated with the Lutetean subdivision of the Eocene of Europe. T. A. Conrad[1] as early as 1855 reported certain described Eocene species he had found in a bowlder obtained from near Fort Tejon, California, sent to him by W. P. Blake.

Later G. D. Harris in his paper entitled "Correlation of Tejon Deposits with Eocene Stages of Gulf Slope"[2] correlated the Tejon with the lower Claiborne on the basis of the identity of highly

[1] T. A. Conrad, "Paleontology," *Pac. R.R. Rept. App. to Preliminary Geol. Rept. of W. P. Blake* (1855), pp. 5–20.

[2] G. D. Harris, *Science*, Vol. XXII (Aug. 12, 1893), p. 97.

ornamented species common to the two, and also because of their generic assemblage.

Dickerson's[1] conclusion was the same as that of Harris. He listed a much larger number of identical or nearly identical species common to the Claiborne and the Tejon.

The nonconformity of the Meganos beds below the Tejon, together with the fact that the faunas of the two groups are very different, would seem to show that the former belong to a horizon lower than that of the lower Claiborne. That it does not represent the lowest Eocene is shown by the fact that the beds of the Martinez Group, which contains a fauna very distinct from that of the Meganos, lie stratigraphically and unconformably below those of the Meganos.

Dickerson's[2] conclusion after studying the fauna of the Martinez was that it is "in part the correlative of the Midway of the Gulf States and in part represents a division of time earlier than the Midway."

From our present knowledge of the Meganos fauna, its relationship appears to be closer to that of the Tejon than to that of the Martinez. If this be true, it would seem improbable that the Meganos is the equivalent of any part of the typical Midway stage. It more probably corresponds to the Wilcox. There is some direct evidence which appears to favor this assumption. This is the presence of species in the Meganos identical or nearly identical with certain well-known Middle Eocene species. For example *Turritella merriami* Dickerson with its numerous variations appears to be specifically close to *Turritella humerosa* Conrad, which is common in the Middle Eocene Wilcox of the Gulf province. One of the new species of turritella associated with *Turritella merriami* in the Mount Diablo region appears to be identical in at least one of its variations with the species listed by Harris[3] as *T. humerosa var.*

[1] R. E. Dickerson, "Stratigraphy and Fauna of the Tejon Eocene of California," *Univ. Cal. Pub. Bull. Dept. Geol.*, Vol. IX (1919), No. 17, p. 476.

[2] R. E. Dickerson, "Fauna of the Martinez Eocene of California," *Univ. Cal. Pub. Dept. Geol.*, Vol. VIII (1914), No. 6, p. 120–21.

[3] G. D. Harris, "The Midway Stage," *Bull. Amer. Paleontology*, No. 1, Pl. 11, Fig. 12 (1896).

A number of the undescribed species from the Meganos of different localities appear to be closely related to Wilcox species, if not identical with them. The correlation of the Meganos Group of California with the Wilcox of the east is only tentatively proposed. It is very possible that when the faunas of the Meganos and the Tejon are better known our ideas as to the relative position of the different Eocene horizons on the West Coast will be considerably revised.

SUMMARY OF CONCLUSIONS

The Meganos Group, the newly recognized division in the Eocene of California, has a wide distribution throughout the Coast ranges of the state. These beds represent an epoch of deposition distinct from both the Martinez below and the Tejon above. At the end of Martinez times there were orogenic movements which caused the sea to be withdrawn from what is now the Coast range province. When the sea again came, during the Meganos epoch, into this general region, its area was considerably different from that of the previous epoch. Following the deposition of the Meganos deposits, crustal movements again caused the withdrawal of the sea from the Coast range region. These movements were general throughout the Coast ranges, and over wide areas the Meganos beds were folded, the deposits of the next epoch of deposition, the Tejon, being laid across the upturned and eroded edges of these folded beds.

The Meganos of the Coast ranges belongs to the same epoch of deposition as the beds of the Siphonalia sutterensis zone, described and referred to as a part of the Ione formation by Dickerson. Previously the Siphonalia sutterensis beds had been considered to represent the uppermost Eocene of the West Coast.

A fairly large well-preserved fauna has been obtained from the Meganos horizon. This is very distinct from that of the Martinez (Lower Eocene). A few of the Meganos species are found in the Tejon. The fauna appears to be more closely related to the Tejon than to the Martinez, and is correlated provisionally with the Wilcox horizon of the Gulf and Atlantic Coast provinces.

VULCANISM AND MOUNTAIN-MAKING: A SUPPLEMENTARY NOTE

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In a recent paper on "The Building of the Colorado Rockies,"[1] the writer compared and contrasted the Rocky Mountains of Colorado, an example of a thick-shelled range, with the Pennsylvania Appalachians, which belong to the thin-shelled type. One of the striking differences developed was that the building of the more deeply rooted Colorado mountains was accompanied by the extrusion of much lava, while the shallower Appalachians were formed with the extrusion of but little lava. A further comparison was made between different portions of the Rocky Mountain chain itself. In Colorado, where the deformed zone extended deep below the surface and the vertical element in the deformation was distinctly large, lava flows appeared in abundance; while in the Canadian Rockies, where the deformed zone was much shallower and suffered intense horizontal thrust, but very little volcanic activity occurred.

The study was really concerned with only these ranges, but some of the ideas developed were extended, as tentative suggestions, to other mountain systems. The Alps, the Scandinavian chain, the Scottish Highlands, and the Serra do Espinhaço of Brazil were cited as probably belonging to the thin-shelled type because of their sharp folding, extensive thrust faulting, and great crustal shortening, and it was stated that "only moderate igneous activity" was associated with their development.[2] As representatives of the

[1] *Jour. Geol.*, Vol. XXVII (1919), pp. 248–51.

[2] This refers only to the strongly deformed portions of these ranges, since any igneous manifestations outside of these belts would also be outside the thin-shelled tracts, even though they be more or less related genetically to the mountain-making stresses. In the case of the Alps, the statement refers of course to the intense Tertiary diastrophism. Whatever relation the granites in the axial portions of the Alpine ranges may have had to the Hercynian and earlier orogenic movements is not here considered.

thicker-shell type of mountains in which vertical movements are more pronounced and horizontal thrusting and shortening less conspicuous, the Tertiary Cascades of the Pacific Coast, the Western Andes, and the Abyssinian Mountains were cited. In contrast with the preceding, the growth of these ranges was marked by the extravasation of vast floods of lava.

In this treatment of the subject it was not explicitly stated that extrusive vulcanism alone was considered, although the text would seem to the writer to convey that idea clearly enough. In view of the fact that intrusions did not enter vitally into the Appalachian section of Pennsylvania or that of the Colorado Rockies,[1] this phase of vulcanism had no place in those studies, and so the topic of intrusions was not introduced into the comparison of thin-shelled and thick-shelled ranges in general. But there are, however, certain cases in which plutonic rocks appear in ranges of the thin-shelled type in such a way as to suggest that intrusions on a large scale may be a common habit of this type. It is because of the feeling that the absence of any statement covering the intrusive phase of vulcanism might convey erroneous ideas, and perhaps lead to more or less justified criticism, that the present note is added.

Many folded mountain ranges of both thin-shelled and thick-shelled types are characterized by cores of crystalline rock, in considerable part of igneous origin. In many of these the crystalline rock clearly belongs to an old terrane arched up in the folding process and exposed by erosion; but in many other cases the intrusive relations of the igneous rock lead to the belief that it was intruded into the axis of the folded range in a late stage of the arching process. The wide prevalence of this phenomenon has been emphasized by Daly in the following terms:

> Granitic intrusion of the batholithic order, to observed levels, always follows periods of the more intense orogenic movement. This implies that the greatest abyssal injections of the earth's crust by magma are genetically associated with the horizontal shearing of a superficial earth-shell which is much thinner than the whole crust.[2]

[1] Except in the pre-Cambrian complex, which has nothing to do with the Rocky Mountain diastrophism.

[2] Reginald A. Daly, "Geology of the North American Cordillera at the Forty-ninth Parallel," *Geol. Surv. of Canada, Mem. 38,* Part II (1912), p. 573.

Cores of granodiorite and allied rocks having intrusive relations with the adjacent sedimentaries are very conspicuous in the sharply compressed Sierra Nevada Range, the Mesozoic Cascades, the Coast Range of British Columbia, and various members of the Cordilleran chain extending into Alaska—a group folded, according to present information, at about the close of the Jurassic. Throughout this disturbed region the Jurasside batholiths are a dominant feature.

In some of these cases, at least, the intrusions seem to follow in consequence of the folding, and to appear beneath the most strongly arched portions of the ranges, presumably in consequence of reduced pressure. That the arching process tends to relieve the pressure beneath and hence favors liquefaction and the penetration of magmas, is a principle long recognized and variously utilized in explaining vulcanism.[1] But whether the arching and partial relief of pressure be a major or a minor factor in the actual liquefaction of the magma, the shearing movements involved in this type of deformation might well facilitate the transfer of magma, and would favor its insinuation near the surface, either as irregular batholithic bodies, or perhaps more likely as large lenticular or pancake-shaped intrusive masses whose thicknesses are much less than their horizontal extent. The general laccolithic shape, using the term in its broadest sense, would seem to be the favorite form.[2] Intrusions of this sort occur in both thin-shelled and thick-shelled ranges.

THIN-SHELLED TYPE

In those thin-shelled ranges in which overthrust faulting has been a dominant feature, intrusions formed in this way should not be conspicuous in the marginal portions where the phenomena of overthrusting are best displayed and the shell was thinnest, but rather in the heart of the deformed belt, where the shell involved in

[1] W. Hopkins, "Researches in Physical Geology," *Phil. Trans.*, Part I (1842), pp. 43–55; Eduard Suess, *The Face of the Earth*, Vol. I (1904), p. 170; W. H. Hobbs, "Some Considerations Concerning the Place and Origin of Lava Maculae," *Beiträge zur Geophysik*, Vol. XII (1913), pp. 329–61.

[2] W. C. Broegger, *Eruptivgesteine des Kristianiagebietes*, II (1895), pp. 116–53; Alfred Harker, *The Natural History of Igneous Rocks* (1909), pp. 60–87.

the diastrophism went somewhat deeper and lifting was relatively more important. As a part of the Jurasside orogeny which developed the Sierra Nevada Range, powerful overthrusting occurred far to the east of these mountains.[1] But in these areas of overthrusting igneous activity contemporaneous with the diastrophism seems to have been unimportant.

The extraordinary Caledonian diastrophism affected both the British Isles and Scandinavia. In the Scottish Highlands on the western border the planes of overthrusting dip eastward under the deformed belt; in Scandinavia they dip westward likewise beneath the strongly deformed belt. Singly, each case illustrates the principle of bordering thrust faults on the outer margins of compressed mountain ranges.[2] Together, Scottish thrusts on the west and Scandinavian thrusts on the east, they constitute seemingly a wedge similar to the Appalachian wedge of Pennsylvania.[3]

The region of these extraordinary Caledonian overthrusts in the Northwest Highlands of Scotland seems to have been essentially free from igneous phenomena during the time of the vigorous deformation.[4] Such intrusions as took place at this time were located off to the southeast, especially in the Ochil and Sidlaw Hills[5] and the Cheviot district, which are near the middle of the deformed belt far from the overthrust border. During the deposition of the Lower Old Red Sandstone which followed the Caledonian disturbance, large quantities of volcanics were poured out in the central Lowlands between the base of the Highland Mountains and the Uplands of the southern counties.[6] But no undoubted vents of Lower Old Red Sandstone age have been detected

[1] C. R. Longwell, "Geology of the Muddy Mountains, Nevada, with a Section to the Grand Wash Cliffs in Western Arizona," *Am. Jour. Sci.*, Fifth Series, Vol. I (1921), pp. 39–62; E. S. Bastin, personal communication.

[2] "The Building of the Colorado Rockies," *Jour. of Geol.*, Vol. XXVIII (1910), pp. 243, 249.

[3] Rollin T. Chamberlin, "The Appalachian Folds of Central Pennsylvania," *Jour. of Geol.*, Vol. XVIII (1910), pp. 228–51.

[4] "The Geological Structure of the Northwest Highlands of Scotland," *Mem. Geol. Surv. of Great Britain* (1907).

[5] Sir Archibald Geikie, *Ancient Volcanoes of Great Britain*, Vol. I (1897), pp. 277–79.

[6] *Ibid.*, pp. 271–72; 295–335.

among either the Highlands on the one hand or the Silurian Uplands on the other.[1]

In Scandinavia the deformed igneous masses resting upon the Cambro-Silurian sedimentaries in the Caledonian mountain belt have been regarded by Törnebohm as portions of the Archean brought from the west by the overthrusting process.[2] If so, contemporaneous intrusions played little part in the overthrust sheets. According to Holtedahl, however, the gneisses are to be regarded as highly pressed younger intrusive masses which, during the deformation, broke forth and moved under enormous pressure from the central belt outward.[3] If in truth these be intrusions related to the Caledonian diastrophism, they are in any case more characteristic of the central belt than of the outer borders.

Among the intensely deformed Cenozoic Alps intrusions of Tertiary age are practically wanting in the central and northern ranges which together make up the region of the *nappes de charriage*, or great overthrust sheets. But in the root region of the Lepontine sheet and the Dinaric zone on the south side of the Alps, from which the overthrust masses are thought to have come, the last phase of strong mountain-building was characterized at various points by intrusions of a granitic nature.[4] Steinmann has already emphasized this contrast between the region of the roots and the region of the sheets.

In addition, it is of course to be noted that quite outside of the true mountainous belt, particularly opposite the inner curves of the arcuate chain (in Hungary, Italy, etc.), volcanic phenomena attained considerable prominence.[5] But the more or less contemporaneous extra-montane vulcanism, though related to the mountain-building stresses, is not a part of this discussion.

[1] Sir Archibald Geikie, *Ancient Volcanoes of Great Britain*, Vol. I (1897), p. 272.

[2] A. E. Törnebohm, "Grunddragen af det Centrala Skandinaviens Berbyggnad," *Kongl. Svenska Vet.-Akad. Handl.*, Vol. XXVIII (Stockholm, 1896), pp. 1–210.

[3] Olaf Holtedahl, "Paleogeography and Diastrophism in the Atlantic-Arctic Region during Paleozoic Time," *Am. Jour. Sci.*, Vol. XLIX (1920), pp. 1–25.

[4] G. Steinmann, "Die Bedeutung der jüngeren Granite in den Alpen," *Hauptversammlung der geol. Vereinigung*, Frankfort (1913), pp. 1–4.

[5] Marcel Bertrand, "Sur la distribution géographique des roches éruptives en Europe," *Bull. soc. géol. de France*, 3e sér., Vol. XVI (1887–88), pp. 573–617; Alfred Harker, *op. cit.*, pp. 20–22, 42.

The last thin-shelled range listed in the paper on the Colorado Rockies was the Serra do Espinhaço of Brazil. In the thrust-faulted portion of this ancient system very little igneous activity of any sort has occurred.[1] Eastward for 170 miles toward the Atlantic Coast, in which strip the heart of this mountain system presumably lay, there remains today only what has been called the Archean Complex. This region is characterized by many massive intrusions. Some of these may possibly have been injected at the time of the Serra do Espinhaço orogeny, though there is no evidence as yet bearing on this question.

Similarly in the Canadian Rockies very little igneous activity occurred in the overthrust region of Alberta; but farther west some of the massive intrusions in British Columbia may prove to have been related to this thrusting.

The outer marginal portions of ranges of this type, both folded and faulted, are particularly superficial. In the great overthrusts but very shallow flakes have been moved. The very low inclination of the fault planes does not carry them to great depths. Beneath the planes of overthrusting, the underlying strata, if of incompetent material, are found to be contorted in many places, but this folding rapidly dies out away from the thrust planes. Such shallow deformation does not greatly facilitate the movement of magmas. But back in the heart of the deformed belt the disturbance goes much deeper, and uplifting with relief of pressure beneath is more pronounced. Here, as the above-mentioned illustrations seem to show, intrusions tend to develop.

THICK-SHELLED TYPE

As pointed out, the thick-shelled mountains have been characterized by open, gentle folding, moderate crustal shortening affecting a deeper zone, by strong uplifting, and the extravasation of much lava.[2] Vertical diastrophism seems to dominate over horizontal. Normal faulting is an important accompaniment, occurring either incidentally as a part of the uplifting process or as a result of subsequent relaxational movements of the raised plateau-like area. Iddings has given an excellent

[1] E. C. Harder and R. T. Chamberlin, "The Geology of Central Minas Geraes, Brazil," *Jour. Geol.*, Vol. XXIII (1915), pp. 341–78.

[2] *Jour. Geol.*, Vol. XXVII (1919), p. 251.

exposition of the part block faulting has played in the extravasation of lava.[1] According to his belief, block faulting under tensile stress offers the principal outlets for the escape of lava. To quote: "The deepest fractures starting from the zone of potential magma should permit its eruption and intrusion between blocks that tend to part from one another by reason of the tensile stress." The wide prevalence of normal faulting in mountains of the thick-shelled type should therefore be an important factor in the rise of magmas. The steep inclination of normal fault planes carries these fractures to greater depths than the more gently inclined thrust faults. At the same time normal fault planes because of the governing tensile stress, at least locally, become the more ready avenues of escape for the lavas. While rhyolite and other acidic lavas have appeared in vast quantities in some places, andesitic and basaltic lavas appear to have been, on the whole, more abundant. This may perhaps be in part because the greater liquidity of the basic lavas makes migration along narrow fissures easier for them than for the more viscous silicic magmas.

SUMMARY

The formation of thick-shelled mountains is characterized in general by much volcanic activity. There may also be important intrusives bearing a close relation to the mountain-making stresses. The growth of thin-shelled mountains, on the other hand, is accompanied by very little volcanic activity, at least within the truly mountainous belt. Little igneous activity of any sort is manifested in the marginal and most strongly overthrust portions of thin-shelled ranges; but in the heart of the deformed belts, where there has been more uplifting and the affected zone goes deeper, granitic and other intrusions are a common and probably characteristic feature. It is of course also to be recognized that a region which, in an earlier age, has undergone deformation of the thin-shelled type may, in a later age, after long continued denudation, participate in orogenic movements of the thick-shelled type and so become the scene of volcanic activity on a large scale.

[1] J. P. Iddings, *The Problem of Vulcanism* ("Silliman Memorial Lectures"), Yale University Press (1914), pp. 79–81, 183–84.

SUMMARIES OF PRE-CAMBRIAN LITERATURE OF NORTH AMERICA

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VI. THE CORDILLERA OF THE UNITED STATES

The most notable advances in the study of the pre-Cambrian of the Cordillera are the finding of pre-Cambrian sediments unconformably below the Lower Cambrian of the southern Sierra Nevada of California by Knopf and Kirk; additions to our knowledge of the extent and composition of the belt series; and the epoch-making investigations of the life-record of the Beltian by Walcott.

Bastin and Hill[1] report that the principal rocks underlying Gilpin County, Colorado, and adjacent parts of Clear Creek and Boulder counties are of pre-Cambrian age. The Idaho Springs formation, a quartz biotite schist of sedimentary origin, is really the most important. Sedimentary origin is inferred from the highly aluminous composition of certain phases, apparent bedding, highly quartzose and certain apparently conglomeratic phases, lime silicate phases probably representing metamorphosed limestone, and the lack of evidence of intrusive relations.

Other pre-Cambrian rocks of the area include stocks of granite gneiss, quartz diorite, and hornblendite, and the younger Silver Plume granite. Granite pegmatites of various ages are abundant.

Blackwelder and Crooks[2] state that the Medicine Bow range, west of Laramie, Wyoming, contains one of the most varied sections in the western United States. They include the basal schist gneiss complex above which are more than 25,000 feet of slightly

[1] E. S. Bastin and James M. Hill, "Economic Geology of Gilpin County and Adjacent Parts of Clear Creek and Boulder Counties, Colorado," *U.S. Geol. Surv., Prof. Paper 94* (1917), 379 pp., 23 pls., 79 figs.

[2] E. Blackwelder and H. F. Crooks, "Pre-Cambrian Rocks in the Medicine Bow Mountains of Wyoming," *Geol. Soc. Am. Bull.*, Vol. XXIX (1918), pp. 97–98.

metamorphosed sediments consisting of quartzites, slates, dolomites, lava flows, pyroclastics, and tillites. Detailed studies have been made.

Butler and Loughlin[1] report that the pre-Cambrian rocks of the Cottonwood–American Fork mining region of Utah, southwest of Park City, consists mainly of shallow-water deposits of quartzites, schists, and slates about 11,000 feet thick. They dip steeply to the north. At the top of the section is a conglomeratic layer resembling tillite.

The St. Joe–Clearwater[2] region of Idaho, comprising about 250 square miles, lies about 30 miles to the southeast of the Coeur d'Alene lead-mining district. The Algonkian sediments exposed in this area include mica schists and quartzites which are correlated with those of the Coeur d'Alene district. The base on which they rest is not exposed. The formation of this area which is correlated with the St. Regis of the Coeur d'Alene lacks the purple color assumed to be characteristic in the type locality. The most difficult correlation was that of the beds believed to correspond to the Newland formation, since they are many times thicker than in the Coeur d'Alene district.

The region around Mullan, Idaho, and Saltese, Montana,[3] overlaps the southeast part of the Coeur d'Alene quadrangle. The pre-Cambrian rocks exposed belong to the Algonkian belt series and include the following formations:

Striped Peak formation	Chiefly greenish gray and some purple beds of shale and sandstone with shallow water markings
Newland (Wallace) formation . . . . 5,000± feet	Blue shale of variable thickness—5,000 feet to insignificant. Blue- and white-banded argillite with some greenish beds, with some limestone beds. Chiefly green colored calcareous rocks with argillaceous beds

[1] B. S. Butler and G. F. Loughlin, "A Reconnaissance of the Cottonwood–American Fork Mining Region, Utah," *U.S. Geol. Surv. Bull. 620* (1915), pp. 165–226, 1 pl.

[2] F. C. Calkins and E. L. Jones, Jr., "Geology of the St. Joe–Clearwater Region, Idaho," *U.S. Geol. Surv. Bull. 530* (1913), pp. 75–86, 1 pl.

[3] F. C. Calkins and E. L. Jones, Jr., "Economic Geology of the Region around Mullan, Idaho, and Saltese, Montana," *U.S. Geol. Surv. Bull. 540* (1912), pp. 167–211, 1 pl. (map).

Ravalli group	St. Regis formation, 1,000± feet—Shales and quartzites of prevailingly green and purple tints with shallow water markings
	Revelt quartzite, 1,200 feet—Thick bedded, in part sericitic, but mostly clean quartzite
	Burke formation, 2,000 feet—Fine-grained, light-colored, thin-bedded siliceous quartzites and shales

Crawford and Worcester[1] report that the pre-Cambrian rocks of the Gold Brick district of Gunnison County, Colorado, include mica, cordierite-mica, garnet, amphibole, quartz, granitic, and hornblendic schists of which some are probably altered sediments. Quartzites and pyroxenic quartzites, quartzite conglomerate, andalusite quartz rocks, and epidote rocks also occur. Certain granite and diorite intrusion are probably pre-Cambrian.

Exposures of quartzite and schist or shale[2] belonging to the Algonkian Uncompahgre formation are found in the Piedra Canyon of the San Juan region.

Darton[3] states that granite is the pre-Cambrian rock of Luna County, New Mexico.

Darton[4] reports that the pre-Cambrian rocks of the Deming quadrangle of southern New Mexico consist of granite with subordinate gneissic granite and diorite.

Duncan[5] describes the pre-Cambrian rocks of Harney Peak, South Dakota, as consisting of mica, hornblende, garnet schists intruded by granite and associated with greisen, pegmatite, and quartz veins.

The Philipsburg quadrangle[6] lies in the central western part of Montana. The Belt series is unconformably overlain by the

[1] R. D. Crawford and P. G. Worcester, "Geology and Ore Deposits of the Gold Brick District, Colorado," *Colorado Geol. Surv. Bull. 10* (1916), 116 pp., 9 pls., 4 figs.

[2] Whitman Cross and E. S. Larsen, "Contributions to the Stratigraphy of Southwestern Colorado," *U.S. Geol. Surv. Prof. Paper 90* (1914), pp. 39–50, 1 pl., 2 figs.

[3] N. H. Darton, "Geology and Underground Water of Luna County, New Mexico," *U.S. Geol. Surv. Bull. 618* (1916), 188 pp., 13 pls., 15 figs.

[4] N. H. Darton, "Description of the Deming Quadrangle, New Mexico," *U.S. Geol. Surv. Geol. Atlas, Folio No. 207* (1917), 15 pp., 5 pls. (maps and illus.), 11 figs.

[5] Gordon A. Duncan, "Contribution to the Study of the Pre-Cambrian Rocks of the Harney Peak District of South Dakota," *Trans. Am. Inst. Min. Eng.*, Vol. XLIII (1913), pp. 207–18, 3 figs.

[6] W. E. Emmons and F. C. Calkins, "Geology and Ore Deposits of the Philipsburg Quadrangle, Montana," *U.S. Geol. Sur., Prof. Paper 78* (1913), 271 pp., 17 pls.

Cambrian Flathead quartzite. This relation is expressed by angular disconformity of bedding and by a basal conglomerate. The Belt formations of this area are correlated with those of the Walcott's Belt Mountain section. The only difficulty in this correlation lies in the absence in the Philipsburg district of beds equivalent to the Greyson shale. The Beltian succession of the Philipsburg district is as follows:

Unconformity

Spokane shales—Shale and sandstone, the prevailing in upper portion, color chiefly red, some cracks, and ripple marks, 5,000 feet

Greyson shales—Apparently lacking, may be included in Newland

Newland limestone—Thin-bedded, more or less siliceous and ferruginous passing into shale, generally buff on weathered surface. Shallow water markings in upper part, 4,000 feet

Ravilli quartzite—Gray with some dark bluish and greenish shale, 2,000 feet

Prichard shales—Dark bluish interbedded with sandstone, rusty brown on weather surface, 5,000±feet

Neipart quartzite—Light colored. Base not exposed. 1,000±feet

Finlay[1] reports that the pre-Cambrian rocks of the Colorado Springs quadrangle consist chiefly of the Pikes Peak granite in which are minor inclusions of acid gneisses and schists. Two other granites and some pegmatite and syenite dikes are included in the pre-Cambrian.

Haynes[2] finds the following pre-Cambrian rocks in the vicinity of Three Forks, Montana:

Empire shale—Even-bedded green shales, interlayered with quartzite. Thickness, 800 feet, (?)

Spokane formation—Well-stratified red and green slates interlayered with mud cracked and ripple-marked sandstone. Thickness, 1,650+ feet.

From studies of the Wardner district quartzite, Hershey[3] adds the Cataldo to the base of Calkins' Belt section.

[1] G. I. Finlay, "Description of the Colorado Springs Quadrangle, Colorado," *U.S. Geol. Surv., Folio 203* (1916), 17 pp., 7 pls., 9 figs.

[2] W. P. Haynes, "The Lombard Overthrust and Related Geological Features," *Jour. Geol.*, Vol. XXIV (1916), pp. 269–90, 11 figs.

[3] O. H. Hershey, "The Belt and Pelona Series," *Am. Jour. Sci.*, 4th Ser., Vol. XXXIV (1912), pp. 263–73.

The Belt series,[1] according to Walcott, unconformably underlies the Middle Cambrian Flathead quartzite. In ascending order the members exposed are the Spokane shale, Empire shale, Helena limestone, and Marsh shale. The total thickness is 3,300 feet at Helena, of which the Helena limestone comprises 2,400 feet.

Knopf and Kirk[2] report that pre-Cambrian rocks underlie with marked erosional unconformity the Lower Cambrian of the Inyo range of southern California. The pre-Cambrian system from the top downward consists of the following:

Deep Spring formation—Local sandstone and dolomite beds, 1,600 feet
Reed dolomite—2,000 feet
? Sandstones and thin-bedded, impure dolomites, 2,000 feet (?)

Both Archean and Algonkian[3] are exposed in the Shinumo quadrangle. They are separated by a well-marked unconformity characterized by angular discordance of structures, difference in metamorphism, and intrusives.

The Archean is composed of the Vishnu schists including quartz, mica, and hornblende schists. They are cut by Archean quartz diorite and granite pegmatite intrusives.

The Algonkian section is as follows:

Unconformity

Dox sandstone—Cross-bedded, ripple-marked in part, mud-cracked argillaceous layers. 2,297 feet plus unknown thickness removed by pre-Cambrian erosion

Shinumo quartzite—Hard, compact, cross-bedded sandstone and quartzite, usually of fine and even grain. 1.564 feet

Hakatai shale—Argillaceous red shale grading upward into arenaceous red shale and sandstone. Nearly all beds contain sun cracks and ripple marks. Metamorphosed by a thick diabase sill. 580 feet

[1] Adolph Knopf, "Ore Deposits of the Helena Mining District, Montana," *U.S. Geol. Sur. Bull. 527* (1913), 143 pp., 7 pls., 4 figs.

[2] A. Knopf, "A Geologic Reconnaissance of the Inyo Range and the Eastern Slope of the Southern Sierra Nevada, California," with a section on the stratigraphy of the Inyo range, by Edwin Kirk, *U.S. Geol. Surv. Prof. Paper 110* (1918), 130 pp., 32 pls., 8 figs.

[3] L. F. Noble, "The Shinumo Quadrangle, Grand Canyon District, Arizona," *U.S. Geol. Surv. Bull. 549* (1914), 100 pp., 18 pls. (incl. map in pocket), 1 fig.

Bass limestone—White crystalline limestone alternating with beds of argillaceous and calcareous red shale containing sun cracks, cut by thick diabase sill. 335 feet

Hotauta conglomerate—Arkose conglomerate characterized by lack of sorting and transportation. Rests on an even surface of erosion. 0–6 feet

Unconformity

Detailed study of the Vishnu series is lacking, but it is thought to be equivalent to the Pinal schists of the Bisbee district.

Paige[1] reports that the pre-Cambrian rocks of the Silver City folio of southwestern New Mexico are granites with minor quartzitic and schistose masses, the whole being mapped as a unit.

Patton[2] *et al.* state that the pre-Cambrian rocks in the Alma district of Park County, Colorado, consist mostly of granitic and banded gneisses and acid schists, granites, pegmatites, and aplites.

Ransome[3] presents ten Paleozoic sections ranging from southern to northern Arizona. In each section Cambrian is unconformably above pre-Cambrian. The pre-Cambrian rocks in these sections are as follows:

District	Pre-Cambrian Rocks
Bisbee	Pinal schists (sericitic schist chiefly metamorphosed sediments cut by granite)
Tombstone	Pinal schist and gneissic quartz-mica diorite
Clifton	Pinal schist and granite
Globe and Ray quadrangles	Pinal schist and intrusive granitic rocks
Roosevelt	Granite
Southern part of the Ancha district	Granite
Northern part of the Ancha district	Granite, older pre-Cambrian schists, younger pre-Cambrian quartzite, and conglomerate

[1] Sidney Paige, "Description of the Silver City Quadrangle, New Mexico," *U.S. Geol. Surv., Atlas Folio No. 199* (1916), 19 pp., 4 pls., 17 figs.

[2] H. B. Patton, A. J. Hoskins, and M. G. Butler, "Geology and Ore Deposits of the Alma District, Park County, Colorado," *Colorado State Geol. Surv. Bull. No. 3* (1912), 284 pp., 29 pls., 6 figs.

[3] F. L. Ransome, "Some Paleozoic Sections in Arizona and Their Correlation," *U.S. Geol. Surv. Prof. Paper 98* (1916), pp. 133–66, 8 pls., 4 figs.

District	Pre-Cambrian Rocks
Head of Canyon Creek	Hornblende rock and schist
Jerome	Schist
Grand Canyon	Grand Canyon series, younger pre-Cambrian including over 4,700 feet of conglomerate, limestone, shale, quartzite, and sandstone resting unconformably on schist of older pre-Cambrian

All the pre-Cambrian rocks in the foregoing sections excepting some of the Grand Canyon and northern Ancha sections are older pre-Cambrian.

Richardson[1] finds that in the region of Castle Rock folio lying between Denver and Colorado Springs, only the youngest of the pre-Cambrian rocks of the Front Range, the Pikes Peak granite is exposed.

Richardson[2] finds that the pre-Cambrian rocks of the Van Horn quadrangle of southwestern Texas are unconformable below the Cambrian and they consist of the following:

Millicon formation—Fine red sandstone, cherty limestone, and conglomerate; in northern Carrizo Mountains

Relations concealed

Carrizo formation—Quartzite, slate, and a variety of schists; in southern Carrizo Mountains

Schultz[3] finds that the pre-Cambrian rocks of southeastern Idaho and western Wyoming include schists, granites, gneisses, and igneous rocks cut by dikes of pegmatite and diabase. Some of the schists and gneisses may be of sedimentary origin. The pre-Cambrian area lies in the central and eastern part of the Teton range.

[1] G. B. Richardson, *United States Geological Survey, Castle Rock Folio No. 198* (1915).

[2] G. B. Richardson, "Description of the Van Horn Quadrangle, Texas," *U.S. Geol. Surv. Geol. Atlas, U.S. Van Horn Folio* (*No. 194*) (1914), 9 pp., 5 figs., 3 maps, illustration sheet.

[3] A. R. O. Schultz, "Geologic Reconnaissance for Phosphate and Coal in Southeastern Idaho and Western Wyoming," *U.S. Geol. Surv. Bull. No. 680* (1918), pp. 84, 2 pls., 8 figs.

Smith and Packard[1] state that certain rocks possibly of pre-Cambrian age have been found in Oregon. They include a granodiorite near the head of John Day River and certain amphibolite, hornblendite, mica quartz schists, and talc schists near the California-Oregon boundary.

Smith[2] states that the pre-Cambrian of California comprises schists and gneisses of Inyo, San Bernardino, and Riverside counties. These underlie Olenellus beds. They also include the Pelona schists of San Bernardino County, the Abrams and Salmon schists of Trinity and Siskiyou counties, the South Fork Mountain schists, and certain old schists and gneisses mapped with the Sierra batholith. Similar rocks also occur in the Sierra Madre.

Somers[3] states that the pre-Cambrian rocks of the Burro Copper district of southwestern New Mexico consist of a complex of granites.

Umpleby[4] reports that in a reconnaissance survey of northwestern Custer County, Idaho, he has found highly metamorphosed schists, slates, and quartzites of Algonkian age, which probably represent a part of the Coeur d'Alene section. The sequence in Custer County was not worked out in detail.

Walcott[5] cites evidence for the unconformity of the Cambrian and pre-Cambrian at Helena, Montana. The specific facts referred to are slight angular discordance, in places an erosion surface on the pre-Cambrian formations, and fossil relationship.

Walcott[6] regards the pre-Cambrian Algonkian limestones of the Cordilleran region of North America as owing their origin chiefly

[1] W. D. Smith and E. L. Packard, "The Salient Features of the Geology of Oregon," *Jour. Geol.*, Vol. XXVII (1919), pp. 79–120.

[2] J. P. Smith, "The Geologic Formations of California with Reconnaissance Geologic Map," *Cal. State Min. Bur. Bull. No. 72* (1916), 47 pp., tables.

[3] R. E. Somers, "Geology of the Burro Mountains Copper District, New Mexico," *Am. Inst. Min. Eng. Bull. No. 101* (May, 1915), pp. 957–96, 25 figs.; *Bull. No. 108*, p. 2476.

[4] Joseph B. Umpleby, "Some Ore Deposits in Northwestern Custer County, Idaho," *U.S. Geol. Surv. Bull. 539* (1913), 100 pp., 10 pls., 4 figs.

[5] C. D. Walcott, "Relations between the Cambrian and Pre-Cambrian Formations in the Vicinity of Helena, Montana," *Smithsonian Misc. Coll.*, Vol. LXIX, No. 4 (June 24, 1916), pp. 259–301, 6 pls., 4 figs.

[6] C. D. Walcott, "Pre-Cambrian Algonkian Algal Flora," *Smithsonian Misc. Coll.*, Vol. LXIV, No. 2 (1914), pp. 77–156, pls. 4–23.

to the action of bacteria and algae. Bacterial remains have not been identified in pre-Cambrian rocks, but numerous concretionary forms have been found in the Newland limestone of the Belt series. These forms are similar to the calcareous bodies formed by modern blue-green algae in fresh-water lakes. Chains of silicified cells which resemble the cell chains of modern blue-green algae were also found. The similarities of structure between the pre-Cambrian and modern algoid forms are clearly demonstrated by a series of plates. Eight Algonkian algal forms are described by Walcott as occurring in the Newland limestone. The Greyson shale overlying the Newland limestone has numerous crustacean remains—*Beltina danai* and many annelid trails representing five species. The next overlying formation of the Belt series, the Spokane shales, is credited with one species of algae.

Walcott[1] briefly describes bacteria which he discovered in the Newland limestone, a formation of the Beltian series of Montana.

The pre-Cambrian rocks[2] of the Dillon quadrangle include the Belt series, 3,000 feet thick, which are composed of slates, thin-bedded quartzite, and schists. They are unconformably above a series of schists and gneisses interbedded with limestones 5,000 feet thick which are correlated with the Cherry Creek group.

VII. THE CORDILLERA OF CANADA

Notable advances have been made in the study of the pre-Cambrian of the southern portion of the Rocky Mountain section by Daly, Schofield, and others. The pre-Cambrian in this section consists of two units, an older complex of clastic and chemical sediments, the Priest River terrane, Shuswap series, etc., folded and metamorphosed, and intruded by granites. The stratigraphic separation of this older unit has not been fully accomplished. Unconformably overlying the basal rocks is a thick series of feebly metamorphosed clastic sediments and limestones, the Belt series, having characteristics of terrestrial sediments. They are coarsest and most fragmental in the western part of the section where they

[1] C. D. Walcott, "Discovery of Algonkian Bacteria," *Nat. Acad. Sci.* (1915), pp. 256–57, 3 figs.

[2] A. N. Winchell, "Mining Districts of the Dillon Quadrangle, Montana and Adjacent Areas," *U.S. Geol. Surv. Bull. No. 574* (1914), 191 pp., 8 pls., 16 figs.

are exposed in contact with the older rocks. Different views as to the upper limits of the Belt series have been held. Daly believed that they were conformable with the Cambrian. Schofield reports that he has found an unconformity at the base of the Middle Cambrian and places the top of the Beltian much higher than Daly.

Allan[1] states that between Banff and Golden in the valley of Bow River along the Canadian Pacific Railway the pre-Cambrian section is as follows: Base not exposed. Corral Creek formation, 1,320 feet composed of quartzites and coarse-grained sandstones with interbedded shales. Hector formation, 4,590 feet, gray, green, purple, siliceous shale with interbedded conglomerates. Remains of brachiopod-like shells in certain beds. Disconformable contact with Cambrian above.

The rocks[2] along the international boundary between the Porcupine and Yukon rivers, classified provisionally as pre-Cambrian, lie on the north side of the Yukon and are peripheral to a larger area of these rocks south of the river. They comprise amphibolites, quartzite schists, mica schists, and occasional limestone beds.

According to Daly,[3] the succession of the Rocky Mountains along the forty-ninth parallel from the Clark range on the western margin of the Great Plains to the Selkirk on the one hundred and seventeenth meridian include the pre-Beltian Priest River terrane; the Beltian; Lower, Middle, and Upper Cambrian; and on the western border of the area, later conformable Paleozoic rocks. The only regional unconformity lies between the Beltian and pre-Beltian.

The pre-Beltian Priest River terrane consists of an undetermined thickness of dynamically metamorphosed sediments, notably mica schists, phyllites, quartzites, chlorites, schists, and dolomites whose stratigraphic order is unknown. The thickness exposed may be about 18,000 feet. They outcrop along the eastern base

[1] John E. Allan, *International Geog. Congress, Twelfth Guide Book* (1913), pp. 167–201, maps 2, prints.

[2] DeLorme D. Cairnes, "The Yukon-Alaska International Boundary between Porcupine and Yukon Rivers," *Canada Geol. Surv. Mem. No. 67* (1914), 161 pp., 2 maps (in pocket), 2 figs., 16 pls.

[3] R. E. Daly, *Geology of the North American Cordillera at the Forty-ninth Parallel* (1912), 3 parts, 840 pp., 73 pls., 42 tables, 17 geologic maps.

of the Selkirk Mountain system in parallel bands striking about north 15° east.

The succession overlying the Priest River terrane is sedimentary with the exception of an extensive basal lava flow, the Purcell lava which occurs in the eastern portion. The extent of the latter makes it a good horizon marker. Geographically, this succession is classified from west to east as the Summit, Purcell, Galton, and Lewis series. The Summit series of the Selkirks is the only one whose base is exposed. It unconformably overlies the Priest River terrane. The various members of these series are correlated with reference to their position above or below the Purcell lava, their mineralogical constituents, and general lithological characteristics, special emphasis being laid upon the molar tooth structure of certain carbonate formations due to the weathering of a peculiar mixture of dolomite and limestone, the presence of a certain orthoclase feldspar with a characteristic microperthitic intergrowth, the presence or absence of red iron oxide, etc. In the Selkirks the sediments are chiefly clastic, conglomerates being important at the base. To the eastward, these grade into finer-grained clastics and carbonate. The easternmost part of the series is composed dominantly of carbonates. Red color is also more prominent in the more easterly series especially in their upper portions.

The only fossils found are the species *Beltina danai* which occur in the Altyn limestone in the lower portion of the Lewis series. This limestone and the one underlying it are classed as Beltian.

The Beltian and Cambrian beds are bent into open nearly north and south trending folds which are disturbed by normal faults, most of which trend in the same direction as the folds. The Lewis series has been pushed over the Mesozoic sediments of the Great Plains along a thrust fault dipping westward at a low angle.

In the case of the Lewis series, the Altyn dolomite and the Waterton dolomite below it are referred to the Beltian because of the presence of *Beltina danai* in the Altyn. The basis for the separation of the series into Beltian and Cambrian is not so apparent. The correlation of the beds designated as Cambrian is based chiefly on the lithological similarity of the Siyeh limestone with the

fossiliferous Castle Mountain limestone of McConnell's Castle Mountain–Bow River series on the Canadian Pacific Railway.

Daly bases his correlation of the other formations on their position with respect to the Siyeh and their lithologic resemblance to certain members of McConnell's sections. Willis placed all of the Lewis series in the Beltian since the *Beltina danai* beds of the Altyn formation are conformable with the overlying beds of the series, whereas in the Belt range the Cambrian unconformably overlies the Beltian and is separated from the *Beltina danai* bearing beds by 7,700 feet of sediments. Walcott also found a plane of unconformity at the base of the Fairview sandstone, the lowest Cambrian in the Bow River section.

Daly[1] describes the rocks along the Canadian Pacific Railway, between Golden and Kamloops, British Columbia.

Pre-Cambrian rocks dominate in this section. He classifies them into the Beltian and pre-Beltian or Shuswap. The two are separated by an unconformity. His divisions of the Shuswap or pre-Beltian follows:

	Formation	Approximate Thickness in Feet
	Unconformity with Beltian System	
Intrusive	Batholiths, laccoliths, sills, dykes, and chonoliths of granite, aplite, and pegmatite, generally metamorphosed	
Shuswap series	Adams Lake basic volcanics (with contemporaneous basic intrusives)	10,000+
	Tshinakin limestone-metargillite	3,900
	Bastion schists (phyllites, etc.)	5,000
	Sicamous limestone (representative of Dawson's "Nisconlith" series)	3,200
	Salmon Arm mica schists	1,800
	Chase quartzite	3,000
	Tonkawatla paragneiss (?)	1,500+
Base concealed		
		28,400

The existence of the individual members of this series is certain, but their relative thickness is still in doubt with the exception of the Tshinakin limestone and the Sicamous limestone.

[1] R. A. Daly, "A Geological Reconnaissance between Golden and Kamloops, B.C., along the Canadian Pacific Railway," *Canada Geol. Surv. Mem. No. 68*, 1915, 260 pp., 7 maps, 46 pls., 4 figs.

The pre-Beltian area forms a core surrounded by Beltian and other younger rocks. It is roughly lenticular in outline, its longest direction extending from northwest to southeast. It is about 400 miles in length and 100 miles in width. The quartzites in this series represent true sandstones, probably derived from a granitic area. Daly believes that the intrusions in the pre-Beltian are of a type characteristic only of the early pre-Cambrian and resemble the Laurentian granites of the Canadian Shield. Sills and *lit par lit* injections are characteristic. The metamorphism of the pre-Beltian, he believes, is due almost entirely to the weight of overlying beds and to temperature. The facts on which this conclusion is based are the parallelism of the fissility and the bedding, the almost complete absence of ordinary folds, the low dip of the beds, and the parallel fissility of the dykes with that of the adjacent beds. Metamorphism of this type, Daly believes, is typical of the pre-Cambrian only. He ascribes it to a steep temperature gradient of the earth's crust and to the abundance of mineralizers in the granitic intrusions of that time. The pre-Beltian of this area, he believes, is probably correlated with the Priest River terrane to the south, the only difference between the two being that as yet no granites have been found in the latter. Equivalent rocks also probably occur on the west shore of Coeur d'Alene Lake.

The Beltian system of the area is classified as follows:

COLUMNAR SECTION OF THE BELTIAN SYSTEM IN THE SELKIRK

		Approximate Thickness in Feet
Top, erosion surface		
Glacier Division ("Selkirk series" of Dawson)	Ross, quartzite (in part)	2,500
	Nakimu limestone	350
	Cougar formation (quartzite with metargillitic beds)	10,800
Albert Canyon Division "Nisconlith series" of Dawson	Laurie formation (metargillite, often calcareous; with subordinate interbeds of limestone and quartzite; basal bed, limestone 50 feet thick)	15,000
	Illecillewaet quartzite	1,500
	Moose metargillite	2,150
	Limestone (marble)	170
	Basal quartzite	280
		32,750
Base, unconformity with Shuswap terrane		

The source of this great thickness of sediments, Daly believes, is the older pre-Beltian terrane, the principal evidence for this being the graduation of coarser sediments to finer in going from west to east. Another fact in favor of this is that the quartzites commonly contain fragments of alkaline feldspar. The quartzites, he believes, are partly dune and loess deposits. The limestones, because of their fine grain, he ascribes to direct chemical precipitation. Most of the sediments are well bedded and consist chiefly of quartz, striated feldspar, and clayey material. Frequently, they show ripple marks and mud cracks.

The Roosville, Phillips, and Rateway[1] formations classified by Daly in his forty-ninth parallel report as Middle Cambrian are assigned by Schofield to the Beltian because the Roosville is unconformably overlain by the fossiliferous Middle Cambrian Burton formation. Both the Purcell and the Galton series of Daly are placed in the pre-Cambrian.

The unconformity between the Burton and the Roosville is not shown by discordance of bedding, but by a basal hematite conglomerate of the Burton, the occurrence of other materials in the Burton which are inferred to have been derived from the Roosville, the striking difference in the metamorphism of the Burton as compared with the formations underlying it, and the occurrence of cryptozoan forms in the formations underlying the Burton.

Schofield[2] describes the Cranbrook area of southeastern British Columbia.

The area is in the southern part of the Purcell Mountains and includes about 50 square miles. The Beltian pre-Cambrian rocks underlie most of the area. Schofield's classification of these rocks follows.

[1] S. J. Schofield, "The Pre-Cambrian (Beltian) Rocks of Southeastern British Columbia, and Their Correlation," *Canada Geol. Surv. Mus. Bull. No. 2* (July 3,1914), pp. 79–91, 1 fig. (map).

[2] S. J. Schofield, "Geology of Cranbrook Map-Area, British Columbia," *Canada Geol. Surv. Mem. No. 76* (1915), 245 pp., 1 map, 33 pls., 15 figs.

Cambrian	Erosion: early Cambrian uplift
Pre-Cambrian (Beltian) Purcell series	Gateway formation: (continental deposition), sandstones, sandy argillites, some concretionary siliceous dolomite. Salt casts and ripple marks
	Purcell lava, Purcell sills: intrusion of gabbro accompanied by outpouring of basalt over land surface
	Siyeh formation: (mainly continental, some possibly marine deposition), red, purple, and green mud-cracked argillites, ripple-marked sandstones, some limestones
	Kitchener formation: (continental and possibly marine deposition), calcareous argillites, argillaceous quartzites ripple marked, mud-cracked, some limestones
	Creston formation: shallow water deposition, quartzites, argillaceous quartzites, mud cracks, and ripple marks
	Aldridge formation: argillaceous quartzites, some conglomerates

Reviews

Mineralogy. By Kraus and Hunt. McGraw-Hill, 1920. Pp. xiv+561. \$4.50.

This is in many respects an excellent text, and one that should have wide use. The matter is presented in a simple, direct style and, while abbreviated, is concise and to the point, and should be suitable for classes in first-year mineralogy. The presence of a moderate number of small photographs of distinguished mineralogists together with extremely brief histories is a feature of the book.

Ninety-four pages are devoted to crystallography, thirteen classes being described in detail. This matter is taken largely from the senior author's excellent *Essentials of Crystallography* (1906), and needs no comment here. As in the earlier work, no photographs of crystals are to be found in the part on crystallography. Instead the authors have included numerous photographs of crystal models. This makes the shapes of ideal crystals clear, but as most crystals found in nature are more or less imperfect, some photographs of crystals would have been of value. Under descriptive mineralogy there are numerous photographs of crystals.

One hundred forty-three pages are devoted to the description of 150 minerals. Although largely taken from the senior author's *Descriptive Mineralogy* (1911), the material is abbreviated and is illustrated by numerous well-selected photographs. The arrangement of the minerals is strictly chemical, the authors even going so far as to put several minerals generally grouped with the oxides into separate divisions such as aluminates, ferrites, manganites, and titanates, and, vice versa, zircon is placed with the oxides. Brucite, prehnite, vivianite, and wad are omitted. Varieties, occurrence, associations, important localities, and uses, as well as the more commonly described features of minerals, are given.

The determinative tables for 150 minerals take 169 pages. They are based on physical differences—luster, color, streak, and hardness. One column in the tables describes the mineral associations. Color seems overemphasized, since it is in general less diagnostic than streak or hardness. As an illustration of this point, ten minerals with a wide range of color were selected, and of these brown biotite, red, yellow, or

brown olivine, and colorless tourmaline could not be found under the proper divisions. These minerals, with the colors given, are, to be sure, more or less rare. The tables appear to be excellent for at least the more common varieties of minerals.

The volume also has chapters on the physical and chemical properties of minerals, the polarizing microscope, the formation and occurrence of minerals, qualitative blowpipe methods, gems and precious stones, as well as a classification of minerals according to the elements they contain, giving their uses and statistics of their production. Six pages are devoted to a glossary.

Perhaps one of the most noticeable defects of the book is its entire lack of references, whether in the form of a general bibliography or as footnotes. There are a number of typographical errors, though but few that might lead to confusion were noticed. There are some other errors, not so surely typographical, such as placing wulfenite under the wolframite group, giving tourmaline the formula $H_{20}B_2Si_4O_{21}$, and placing the origin of the minerals in igneous rocks under minerals formed from fusion (rather than from solution). The chemical distinction between the different plagioclase feldspars is poor.

In spite of these minor defects the work is excellent, well arranged, and attractively presented.

D. J. F.

Geology and Mineral Resources of the Hennepin and La Salle Quadrangles. By GILBERT H. CADY. Illinois Geological Survey, Bull. No. 37. Urbana, 1919.

The area represents one of the richest agricultural, manufacturing, and mining communities in the Middle West, including parts of La Salle, Bureau, and Putnam counties of north-central Illinois. The manufacturing wealth is dependent to no small degree upon the natural resources of the region.

General geology.—The area forms a part of the Glaciated Plains Province, the larger part of it being monotonously level, what relief there is being chiefly due to glacial drift. Stratigraphically the rocks of the region range from Cambrian through the Carboniferous, with Pleistocene glacial drift. Good detailed sections and faunal lists are given. There are several important unconformities. Though the strata are for the most part nearly horizontal, the La Salle anticline, or, as the author suggests, more properly the La Salle monocline, shows dips up to 50° locally.

Mineral resources.—The chief resources of the area include coal, limestone for the cement industry, sand and gravel, shale and clay, glass-sand, building stone, and water. Coal is of great importance, the region lying within the northern border of the eastern interior coal basin. While coal is present at several horizons in the Pennsylvanian, only the La Salle bed of the Carbondale formation is at present being mined. Workable deposits of clay and shale are confined to the Pennsylvanian and Quaternary deposits. Both are of some importance. The Lower Magnesian limestone formerly was extensively used as a source of natural cement and is still used to a small extent. One of the important industries of the region is the manufacture of Portland cement. The La Salle limestone is the chief source. The St. Peter sandstone is the chief source of building stone. Sand and gravel are abundant. While the structure is favorable, no evidence of the accumulation of oil and gas has been found other than the local subglacial pockets of gas which are in places utilized by private individuals.

A. C. McF.

The Artesian Waters of North Eastern Illinois. By Carl B. Anderson. Illinois State Geological Survey, Bull. No. 34. 1919.

The importance of this report lies in the partial dependence of the industries and cities of northeastern Illinois on artesian waters. Exclusive of Chicago, whose industries consume 30,000,000 gallons daily, the people of the region depend almost entirely on artesian waters. The area investigated includes thirteen counties.

The bedrock comprises Cambrian, Ordovician, Silurian and Pennsylvanian strata. It was found that while some water is obtained from all the systems named, the Potsdam sandstone is the best available source, a yield of 200 gallons per minute being common from flowing wells. For private consumption the shallower horizons are better. In general the water is hard.

The method of treatment is by counties with more detailed discussions for the different cities and villages. A large amount of useful data are compiled in the report, including water analyses, drilling costs, temperature of the water, and the position of the water table.

A. C. McF.

Geology of Petroleum. By WILLIAM HARVEY EMMONS. First Edition. McGraw-Hill Book Company, 1921.

It is gratifying to record in this book by Professor Emmons a most notable contribution to geological education and one that was imperatively needed, because of the unusual present interest in geology as applied to the occurrence of petroleum and natural gas.

While the reviewer is not entitled by experience to pass judgment on the book from the viewpoint of the advanced specialist in petroleum geology, surely from the standpoint of the general geologist interested in petroleum occurrence the volume supplies a most welcome means of acquiring familiarity with special geologic problems connected with oil occurrence, and, as stated in the preface, the book was designed to meet this need on the part of professional geologists, and especially of college students already familiar with the fundamental principles of geologic science.

The work is based on a series of lectures which for several years have been offered in courses in economic geology at the University of Minnesota. Especially noteworthy is the large number of references indicating that the author has made an exhaustive search of the literature. This feature makes the volume valuable as a source book, in addition to its primary purpose. Two hundred and fifty-four text figures, mostly line drawings, are judiciously selected.

Chapter I, the introduction, includes sections on geographic and geologic distribution. Tabulations of geologic distribution bring out the notable fact that only in the United States have important oil reserves been found in rocks of Paleozoic age. Chapter II deals with surface indications of petroleum and methods for testing rocks for the presence of oil.

Other chapters deal with the openings in rocks, the association of petroleum and salt water, reservoir rocks and covering strata, and the properties of petroleum and natural gas. Current views concerning the first stages of the formation of petroleum are covered in other chapters, as well as theories governing the accumulation of oil in pools. A very useful table is given showing the salient features of a large number of oil fields, including the age of the rocks, their kind, the nature of the cover, the structural character of the reservoirs, and the nature of the surface indications of petroleum.

A long chapter is devoted to the structural features of oil and gas pools, and chapters of lesser importance are devoted to the effects of

dynamic action on petroleum occurrence, gas pressure and oil recovery, and petroliferous provinces.

While a short chapter deals with the subject of maps and logs and explains the meaning of structural contours, there is no detailed discussion of methods of field work in petroleum geology, nor does the book touch the ground of petroleum technology already adequately covered by other works.

The general features of petroleum occurrence mentioned above occupy the first one hundred and ninety-four pages. Nearly three hundred pages are devoted to summaries of petroleum occurrence in the various fields in the United States, and about one hundred and twenty pages to foreign occurrences.

E. S. B.

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PREFACE

The writer's experience in the Adirondack Mountains of New York state has led him to realize the lack of published information concerning the perthitic feldspars. The dominant feldspar of the Algoman augite syenite of the region is perthite (microperthite). This Precambrian rock is a differential phase of an important rock unit which grades from syenites, through quartz syenites, and syenite granites to true granites. In passing from the femic to the salic phases it is readily observed that the feldspar becomes more potassic and passes into what is commonly known as orthoclase, and thence into microcline. Thus the Algoman syenite-granite rocks raise two questions: What is perthite? And what is the difference between orthoclase and microcline?

The attempt to answer these and similar questions led the writer to make the investigation of which the following paper is the result.

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THE MINERALOGRAPHY OF THE FELDSPARS

PART I

HAROLD L. ALLING
Rochester, New York

INTRODUCTION

The rock-forming minerals can be studied (1) as solids capable of assuming definite geometrical forms—crystallography; (2) as optical media that affect light in characteristic ways—determinative petrography; (3) as chemical substances—the chemistry of silicon compounds; and (4) as the end products of crystallization of melts—geophysical chemistry, or as here proposed—mineralography.

This last classification embodies the point of view which has been adopted in this paper for the study of the feldspars, treating them as chemical substances formed by the solidification of melts. When the important mineral groups are fully examined, and their various thermo-equilibrium diagrams published, then the interpretative petrographer will be able to unravel the life-histories of rocks with much greater accuracy and in finer detail than is possible at present.

Appreciating the value of metallographic methods as applied to the elucidation of the feldspar system, it is therefore proposed that these silicates be examined with the aid of the phase rule. The phase rule has shed a flood of light upon the nature and constitution of alloys, and promises to be of equal value in the study of magmas and their crystallization into igneous rocks.

The components here considered are the K-, Na-, Ca-, and Ba-feldspars, which are assumed to be stable under the conditions of the present investigation. There is no difficulty in understanding the term "component" employed here as an example of the nomenclature of the phase rule. It is well for the purpose of this discussion that we understand the term "phase." "Phases are the homogeneous states, whether of freedom, solution, or combination, and whether solid, liquid, or gaseous, into which the components present pass or group themselves . . . . the phases are the transitory stages, states, or conditions, physical or chemical, through which the components pass as they are heated up and cooled down, or as their pressure rises and falls."[1]

The mathematics of the phase rule applies to systems which are in a state of perfect equilibrium. It cannot be disputed that many metallic alloys and silicate complexes (silicate alloys) are in a state of imperfect or false equilibrium. So frequently does this condition occur that the phase-rule method of considering such systems has been the object of considerable criticism. These critics are entirely justified so far as the strict application of the rule is concerned, but apparently overlook the value of thermo-equilibrium diagrams. The use of the diagrams and the application of the phase rule are distinct. Such diagrams help to explain the phenomena of crystallization and microscopic textures of silicate and metallic alloys. Howe[2] points out that hardened steel is a metastable system, that is, it is not in equilibrium, and yet that fact does not depreciate the value of the iron-carbon diagram to the steel manufacturer. A condition of imperfect equilibrium can be definitely indicated upon the diagram. Ostwald[3] says: "In spite of my great admiration for the progress that has been made in metallography through the introduction into this field of the concepts of chemical equilibrium, the phase rule and van't Hoff's concept of solid solution, I cannot help emphasizing the need of caution in all this, for those concepts are all based upon the truth

[1] H. M. Howe, *The Metallography of Steel and Cast Iron*, 1916, p. 240.

[2] *Ibid.*, p. 231.

[3] Wolfgang Oswald, *Theoretical and Applied Colloid Chemistry*, translated by Martin H. Fischer (1917), p. 198.

of certain assumptions. . . . . The belief that true equilibria are attained in these solid mixtures . . . . lacks support."

This quotation is quite typical of such criticisms: it gives the impression that in studying a system, such as the feldspars, by means of the phase rule the laws of physical chemistry are violated. Metallographers such as Howe, for example, are recognizing the metastability of many systems. Many feldspars are metastable solid solutions.

It would appear that the most satisfactory method of attack would be a laboratory study of the melting and freezing phenomena of the feldspars. The work of Allen and Day[1] has shown, however, that little information can be secured through this means because of the high viscosity of the alkali feldspars when above their fusion temperatures. Even when in a molten condition it is impossible to effect crystallization. Watts[2] using pyrometric cones in his work upon artificial mixtures of natural potash-soda feldspars has found it very difficult if not impossible to secure reliable thermal data.

Thus the method of attack here is in part a reverse process. By a study of textures and properties of various specimens differing in their chemical composition, the thermo-equilibrium diagrams of the binary systems can be inferred by analogy with the diagrams of other systems and by means of suggestions found in the metallurgical and mineralogical literature. The writer suggests that the application of the methods of the metallographer to the study of silicate systems for the purpose of increasing our knowledge concerning them may properly be called the science of mineralography.

Murdoch[3] has already proposed the use of this term in a slightly different sense to cover the employment of the metallurgical (metallographic) microscope in the study of non-transparent minerals in the same way that it has been used for years in the study of metals.

[1] E. T. Allen and A. L. Day, "The Isomorphism and Thermal Properties of the [Plagioclase] Feldspars," *Carnegie Inst. of Wash. Pub. 31*, 1905.

[2] A. S. Watts, "The Feldspars of the New England and North Appalachian States," *U.S. Bur. Mines Bull. 92*, 1916.

[3] Joseph Murdoch, *Microscopial Determination of the Opaque Minerals*, 1916, p. iii.

It seems to the writer that the analogy is not perfect. Metallography deals with the constitution, crystallization, and microscopic textures of metallic alloys and their relation to their physical properties. Such study must necessarily be guided by constant references to the thermo-equilibrium diagram of the system under consideration. In the examination of opaque minerals, such as the sulphide ores, the diagrams are not at hand, and not until then should the term mineralography be employed. Furthermore it is questionable whether it is best to limit mineralography to the opaques as contrasted with the transparent minerals. Mineralogy and petrography become mineralographic when the phase rule is considered and the interpretation of mineral compositions and rock textures studied in the light it sheds upon them. This paper is an attempt to discuss the essential nature and relationships of the feldspars, together with an explanation of the methods used in studying them. Discussions of some phases of this subject have appeared in German, but no adequate study of the entire system has been attempted in this country. The following brief summary indicates the substance of the discussion, though not the precise order of presentation.

Many of the principles that apply to metallography and metals apply also to minerals. One of these essential principles is that of solid solutions. To handle this principle the method known to physical chemists as the phase rule has been applied to metallography, necessitating the construction and interpretation of thermo-equilibrium diagrams to set forth the constituents and the processes which form metallic alloys.

Now the principle of solid solutions is applicable also to minerals, and is indispensable to a thorough knowledge of their composition. The mineralogist and the geologist should therefore avail themselves of the phase-rule method, and should understand the use of its diagrams if they would thoroughly grasp the nature of the constituents and of the physical phenomena which have given them their rocks and minerals, the end products of such processes.

This study involves: (1) the construction of thermo-equilibrium diagrams; (2) the interpretation of such diagrams; (3) the application of these diagrams to the rock-forming minerals.

The present-day conception of the chemical nature of the silicates is the result of slow development. In 1846 Laurent[1] suggested that the silicates should be considered to be salts of several acids rather than a single one, and by 1865 ortho-, meta-, and trisilicic acid had become a firmly established nomenclature. Later Vernadsky[2] pointed out that in some aluminum-bearing silicates the element aluminum seemed to possess the characteristics of an acid. Thus the theory that some silicates are aluminosilicates instead of simple silicates was developed and is entertained by many geochemists. While it would be instructive to follow this phase of the subject in greater detail it is outside of the main purpose of this paper. However certain aspects of the aluminosilicate theory need to be considered in attacking the problem of the relationships between the soda, potash, and lime feldspars. "Schwantke[3] while reasoning over the rôle of the lime-silicate which occurs mixed with the potash silicate $K_2Al_2Si_6O_{16}$ in orthoclase built up theoretical conclusions of much interest. . . . . In order to make the formulas of albite, anorthite, and orthoclase more analogous to each other, we may write them $Na_2Al_2Si_2Si_4O_{16}$, $K_2Al_2Si_2Si_4O_{16}$, $Ca_2Al_2Al_2Si_4O_{16}$, and $Ba_2Al_2Al_2Si_4O_{16}$."[4]

Bayley[5] says that chemically, the feldspars may be regarded as isomorphous mixtures of the four compounds, $KSiAlSi_2O_8$, $NaSiAlSi_2O_8$, $CaAlAlSi_2O_8$, and $BaAlAlSi_2O_8$, each of which except the fourth has been found nearly pure in nature. . . . . The feldspars have also been regarded as salts of the acid $H_5AlSi_2O_8$ in which the hydrogen is replaced by various radicals, thus $(KSi)AlSi_2O_8$, orthoclase; $(NaSi)AlSi_2O_8$, albite: $(CaAl)AlSi_2O_8$, anorthite; and $(BaAl)AlSi_2O_8$, celsian. There are several objections to Bayley's conceptions. In the first place, as will be pointed out later, the

[1] Laurent, "Sur les Silicates," *Comp. Rend.*, XXIII (1846), 1050–58.

[2] W. Vernadsky, "Über die Sillimanitgruppe und die Rolle des Aluminums in den Silicaten," *Bull. d. Russ. Ges. d. Naturf.*, 1891, nr. 1–100. (In Russian.)

[3] A. Schwantke, "Die Beimischung von Ca in Kalifeldspath und die Myrmekitbildung," *Centralbl. für min. Geol. und Pal.* (1909), pp. 311–18.

[4] J. J. Serderholm, "On Synantectic Minerals," *Bull. de la commo. geól. de Finlande No. 48*, 1916, pp. 90–91.

[5] William S. Bayley, *Descriptive Mineralogy*, Appleton (1917), p. 408.

feldspars do not form a complete isomorphous series. Secondly, it is not customary to regard silicon as a base-forming element.

(CRYSTALLINE) SOLID SOLUTIONS

The whole subject of solid solutions was opened up in a paper by van't Hoff[1] which appeared in 1890, bearing the title of "Solid Solutions." He said: "If we regard a solid solution as a solid, homogeneous mixture of several substances, the composition of which can be changed without destroying the homogeneity, analogous to solutions in liquids as the solvent, it should not be difficult to cite cases which belong unconditionally in this category." This definition of solid solutions gives us a concept that is a most valuable aid to the study of substances in the solid state.

Again van't Hoff defines a solution as "a homogeneous mixture, the composition of which can undergo continuous variation within the limits of its stable existence,"[2] or in other words a solid solution is a solid homogeneous complex of several substances, whose proportions may vary without loss of homogeneity. A solution may be defined in terms of the phase rule as "a homogeneous mixture which undergoes a change in composition in producing a new phase."[3] Washburn says that "a solution may be defined as a one-phase system composed of two or more molecular species."[4]

Ordinarily the term "solution" has been limited to solids dissolved in liquids. But recent investigation has shown that the conception of solutions should be extended to include gases and solids as actual solvents. The metallographer and the geologist are beginning to speak of "solid solutions," a term which has cleared the atmosphere and opened the way toward a better understanding of matter in the solid state.

The terms "solid solution" and "crystalline solid solution" have both been used in referring to solids dissolved in solids. But the latter term is not as common, therefore the term "solid solution" will be used in this discussion.

[1] J. H. van't Hoff, "Solid Solutions," *Zeitschr. phys. Chem.*, V (1890), p. 322.

[2] As cited by J. V. Elsden, *Principles of Chemical Geology*, 1910, p. 116.

[3] J. L. R. Morgan, *The Elements of Physical Chemistry*, 1918, p. 147.

[4] Edward W. Washburn, *Principles of Physical Chemistry*, McGraw-Hill, 1915.

Solid solutions are frequently referred to as "mixed crystals," a term probably derived from a too literal translation of the German "Mischkrystalle." This translation is unfortunate since the term may suggest a merely mechanical mixture of crystalline substances like sand. As the term is liable to convey an entirely erroneous impression, it should be avoided in the interest of clearness.

> Owing to the restricted nature of molecular motion in crystals and the high viscosity of this state of aggregation, crystalline solutions in a state of equilibrium are seldom met with in practice, because the attainment of equilibrium in a reasonable length of time is so frequently prevented by the restraints upon the free movements of the molecules.[1]

This fact prevents satisfactory laboratory work on the problem. In geology we have at our disposal enormous lengths of time in which perfect equilibrium may be established, a factor, the importance of which is not always appreciated and which of course is rarely attained in the laboratory.

Although the writer is inclined to the view that the feldspars are aluminosilicates which in many cases are capable of forming solid solutions, nevertheless the usual nomenclature (ortho-, meta-, and trisilicates) will be employed without necessarily implying any particular theory of the chemical nature of these minerals.

ISOMORPHISM AND SOLID SOLUTION

Recognizing that isomorphous substances must be completely soluble in each other some mineralogists have come to the conclusion that the terms "isomorphism" and "solid solution" are synonymous. As this matter is of some importance here it will be discussed.

"The establishment of complete solid solution between the [plagioclase] feldspars raises the whole question of the use of the terms solid solution and isomorphism. Some authors use isomorphism to designate complete solid solution, others speak freely of limited isomorphism, and still others use the term in its original significance of simple crystallographic similarity."[2]

[1] E. W. Washburn, *Principles of Physical Chemistry*, 1915, p. 117.

[2] N. L. Bowen, *Amer. Jour. Sci.* (4), XXXV (1913), 595.

Hlawatsch,[1] in a very complete review, concludes that two minerals are isomorphous when: (1) they possess like crystal forms; (2) when their chemical composition is strictly analogous; (3) when they are capable of forming homogeneous "mixed crystals." The etymology of the term isomorphism strictly means "same form." We know, however, that two minerals that are isomorphous do not possess identically the same crystal form but differ from one another in many details. Though they are similar they are not identical. In addition to the crystallographic meaning of the term a similarity in chemical composition is implied. On the other hand in viewing such substances mineralographically the term is applied to systems where complete solubility in the solid condition exists among the components.

One of the difficulties in the correct use of these terms is illustrated by their application to the plagioclase feldspars. Albite is soluble in all proportions in anorthite; both of these minerals are triclinic but their formulas as generally written indicate respectively a trisilicate and an orthosilicate. Their mutual solubility has suggested to mineralogists that their chemical structures ought to be similar, being similar salts of the same acid as suggested by Schwantke[2] and Bayley[3]

But this gets us into an opposite difficulty. If the different salts are salts of the same acid, and have similar structures, how is it then that certain pairs are completely soluble while others are only partially so? To this question no direct answer is forthcoming, for "very little is yet known about the physical and chemical conditions which determine the solubility of a substance in a (liquid) solvent. In fact the essential nature of the process of solution must be regarded as at present uncertain. It has been noticed that solution is more likely to occur if the [components] are alike chemically, but no general rule can be framed."[4] On the other hand when we consider solids and (crystalline) solid

[1] Hlawatsch, *Zeitschr. für Kryst.*, Vol. LI (1912), pp. 417–91.

[2] A. Schwantke, "Die Beimischung von 'Ca' in Kalifeldspath und die Myrmekitbildung," *Centralbl. für min. Geol. und Pal.*, 1909, pp. 311–18.

[3] William S. Bayley, *Descriptive Mineralogy*, Appleton, 1917, 408.

[4] W. C. D. Whetham, *Theory of Solution*, Cambridge University Press (1902), p. 78.

solutions we find a more satisfactory answer to our query. Some fine work on this subject has been done by the Italians, especially by Ciamician and his co-workers.

Ciamician[1] and Bruni[2] have shown that single-ring organic compounds can form solid solutions only with other single-ring compounds. Thus, benzene (not benzine), Figure 1-(1), can form solid solutions only with such compounds as: thiophene, Figure 1-(2), pyrrol, Figure 1-(3), and pyridine, Figure 1-(4). In the same way a double-ring compound can only form a solid solution with another double-ring compound. Thus naphthalene, Figure 1-(5), only with quinoline, Figure 1-(6), etc. Under that same rule applied to triple-ring compounds anthracene, Figure 1-(7), forms solid solutions with its isomer, phenanthrene; with carbozol, Figure 1-(8), or with other triple-ring structures.

The complete solubility of one solid in another is more likely to occur if the chemical structure of each is similar. The similarity of the two, however, does not necessarily have to be so close as the term "isomorphism" would imply. For example some feldspars contain nephelite in solid solution, but this mineral cannot be regarded as isomorphous with the normal feldspar components. It does not seem unreasonable to assume that the mineralographic term "solid solution" is more comprehensive than the crystallographic term "isomorphism."

"It is convenient to regard . . . . [an isomorphous series] as formed by the replacement of one element or radical by another isomorphous with it, rather than as a mixture of different individual molecules."[3] It certainly is true that the term "replacement" is very commonly used, and no criticism to such use can be made provided that the conception of the solubilities of the components is not lost sight of, but there is danger of confusion in such cases. To the economic geologist the word "replacement" immediately suggests such phrases as "pyrite replaced by chalcocite" or "limestone replaced

[1] Ciamician, *Zeitschr. für Physt. Chem.*, XIII (1894), 1; XVIII (1894), 51; XLIV (1903), 505.

[2] G. Bruni, *Rendiconti dell' Accademia dei Lincei.*, Vol. VIII (1899), p. 570.

[3] A. J. Moses and C. L. Parsons, *Mineralogy, Crystallography, and Blowpipe Analysis*, 5th ed., 1916, p. 234.

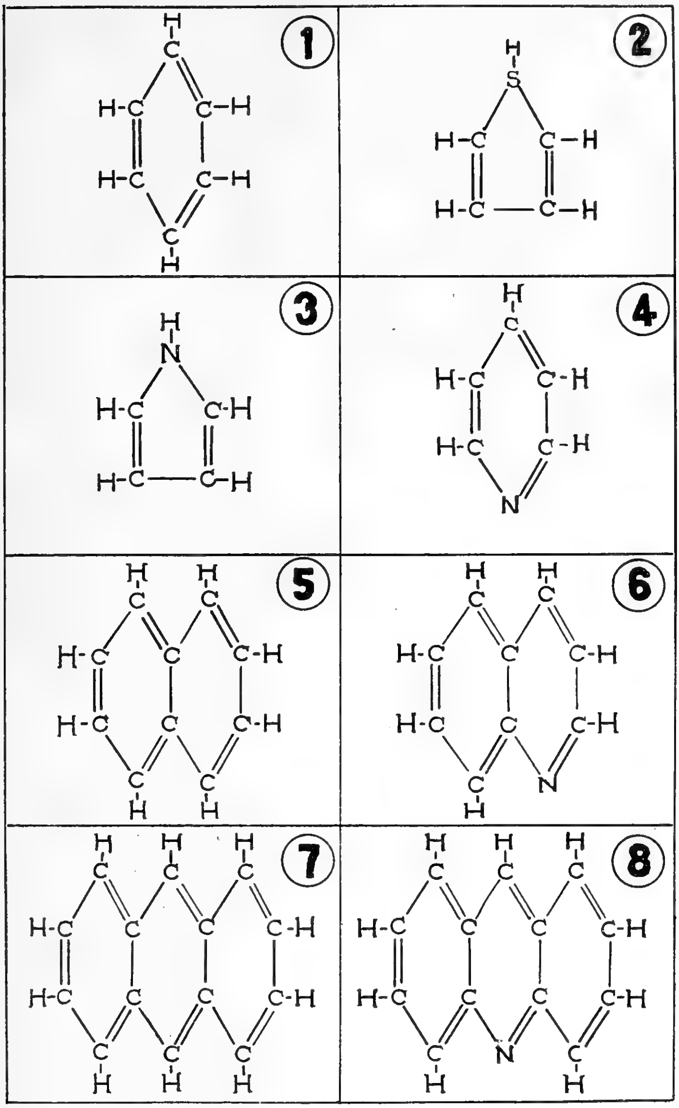


FIG. 1.—Structural chemical formulas of single-, double-, and triple-ring organic compounds, illustrating structures that form solid solutions. 1. Benzene. 2. Thiophene. 3. Pyrrol. 4. Pyridine. 5. Naphthalene. 6. Quinoline. 7. Anthracene. 8. Carbozol.

by ore" due to metasomatism. The term, however, is used in a very different sense by mineralogists. Is it chemically correct to speak of an element in a formula as "partially replaced" or "replaceable" by another? A replacement in a chemical sense means double decomposition taking place through chemical reaction, involving a thermal change; chemical reaction and thermal change being functions of each other. Thus if the potassium of orthoclase is "replaced" by sodium by chemical reaction then there should necessarily be a change in the thermal state of the system, the absorption or liberation of heat. While it cannot be stated positively that there is no change in the thermal state when liquid albite is added to liquid orthoclase at the same temperature, it seems more than likely that the albite will pass into solution without greatly disturbing the thermal equilibrium. The same principle probably applies to solid minerals, although it would be very difficult to prove it experimentally.

Of all the mineralogical systems studied up to the present time none is better known and understood than that of the plagioclase feldspars. It can be asserted with considerable emphasis that the variations in composition in the series are not due to chemical replacement but that they constitute a series of solid solutions. Moreover the plagioclase feldspars are not unique in nature, for many similar systems undoubtedly exist to which the same principle applies.

The reader should not overlook the fact that in this discussion of the feldspars we are dealing with them from the point of view which considers them solid solutions, and mixtures of solid solutions, and not "mixtures, admixtures, and mixed crystals" nor "molecules" which have certain portions "replaceable" by analogous units.

Thus the evidence that can be brought to bear upon this difficult problem points to the conclusion that $NaAlSi_3O_8$, when it exists as albite, and $CaAl_2Si_2O_8$ as anorthite probably possess similar structures. The suggestion of Bayley that the $NaAlSi_3O_8$ in orthoclase is barbierite (see later under "Feldspar Components") may be true but in view of the fact that other evidence points to a possible inversion of albite to barbierite (in the same way that orthoclase

probably changes to microcline) at temperatures below the freezing-point of any melt in the system, it does not seem very likely that the soda feldspar in a magma slowly cooling at great depths could take the form of barbierite. It is very likely that barbierite and microcline, when found, are usually not in their original form, but are the result of inversion, or transition.

The possible inversion of the potash and soda feldspars introduces a new element into the subject. Microscopic evidence does not seem to verify the supposition that microcline and barbierite are high-temperature forms, but rather that they are the result of inversion or that the presence of mineralizers has lowered the temperature of freezing below the inversion range of these minerals.

From the foregoing we conclude that orthoclase and albite have dissimilar structures, while barbierite and orthoclase possess similar atomic groupings. In an analogous way albite and microcline are probably alike.

FELDSPAR COMPONENTS

Modern textbooks on mineralogy list the following feldspars: (1) orthoclase, (2) microcline, (3) soda-orthoclase, (4) soda microcline, (5) anorthoclase, (6) albite, (7) oligoclase, (8) andesine, (9) labradorite, (10) bytownite, (11) anorthite, (12) plagioclase, (13) perthite, (14) celsian, (15) hyalophane, (16) carnegieite, and (17) anemousite. All the feldspars from orthoclase to and including perthite (1 to 13 inclusive) belong to a three-component system where the three units are: (1) potassium feldspar, (2) sodium feldspar, (3) calcium feldspar. Or if celsian, hyalophane, carnegieite, and anemousite be included then all the 17 members fit into a five-component system, of which the following are the components:

Type	Empirical Formula
Potassium Feldspar	$KAlSi_3O_8$
Sodium Feldspar	$NaAlSi_3O_8$
Calcium Feldspar	$CaAl_2Si_2O_8$
Barium Feldspar	$BaAl_2Si_2O_8$
Carnegieite	$Na_2Al_2Si_2O_8$

THE POTASSIUM COMPONENT

Orthoclase.—This mineral (ὄρθος, straight, and κλάς, angle), $KAlSi_3O_8$, is usually regarded as a salt of trisilicic acid, $H_4Si_3O_8$. In regard to its chemical composition *in nature* and its distinction from microcline we shall have more to say.

Microcline.—The name (μικρός, small, and κλίνειν, to incline) has reference to the fact that the angle (89°30′) between the two perfect cleavages differs but little from a right angle. Although the chemical formula of microcline is given as identical with that of orthoclase the petrographer usually has little difficulty in recognizing and separating it from orthoclase by the multiple pericline or "gridiron" twinning of the former as revealed by the microscope. Orthoclase either does not twin in this manner or does so on such an extremely fine scale as to be submicroscopic. Whether there is any physical or chemical difference between the two potash feldspars has been a matter of debate for some time. Investigators are divided between the two general theories; one group maintains that they are one and the same mineral, and differ from one another only in the magnitude of the twinning. On the contrary, another group maintains that there is a fundamental difference between the two minerals which although of identical composition possess different chemical structures.

Dimorphism of the Potassium Component.—"It appears highly probable that if the cross twinning and interpenetration of microcline become so minute as to be invisible under the microscope the crystals would be indistinguishable from those of orthoclase, and would, in fact, possess all the properties of that mineral. Many authors regard orthoclase as pseudo-symmetric; if this be so, all the feldspars may be in reality anorthic [triclinic]."[1]

According to E. Mallard and A. Michel-Lèvy it seems highly probable that orthoclase and microcline are not dimorphous, but identical, since they proved that the optical behavior of orthoclase would be a necessary consequence of an intimate multiple twinning of microcline lamellae after the albite and pericline law.[2]

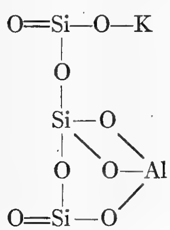
[1] H. A. Miers, *Mineralogy*, p. 460.

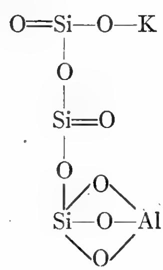
[2] Rosenbusch-Iddings, *Microscopic Physiography of the Rock-Forming Minerals*, p. 320.

On the other hand Vogt[1] speaks of the "probable occurrence of an inversion point in the potassic feldspar from an alpha to a beta modification, viz., from orthoclase to microcline." Barbier[2] has expressed the opinion that potash feldspar exists in two distinct forms which he regards as dimorphous, and which he says is a case of polymorphism. Assuming that there are two dimorphous forms it is not known whether the difference is due to "polymerism" or "isomerism." The term "polymerism" is applied where two compounds possess the same chemical composition but differ in their molecular weights.[3] For example, butyric acid, $C_4H_8O_2$, and aldehyde, C_2H_4O, differ by polymerism. On the other hand "isomerism" is used when two or more compounds possessing the same chemical composition differ in physical or chemical properties but have the same molecular weight, such as levulose and dextrose. If the difference between orthoclase and microcline is due to polymerism then the following formulae might express the relation:

$$(KAlSi_3O_8)n \qquad \text{and} \qquad (KAlSi_3O_8)\ (n+m)$$

where n and m are whole numbers. If the difference is due to isomerism then the kinship of the two minerals may be represented, as suggested by Clarke,[4] as follows:





[1] As cited by C. H. Warren, *Proc. Amer. Acad. Arts. Sci.*, LI, No. 3 (1915), p. 144.

[2] Ph. Barbier, "Researches sur la composition chemique de feldspaths potassique," *Bull. Soc. franç. minéral.*, XXXI (1908), 152–67.

[3] Since the examination of crystals by X-ray spectra has shown that our conception of molecules in solid matter has no real basis, it is well to bear in mind that such expressions as here used are subject to subsequent revision.

[4] F. W. Clarke, *U.S. Geol. Surv. Bull. 588*, p. 12.

The exact nature of the isomeric equivalences among the feldspars is not clear; they may be due to the structure of the salts independently of the acids they represent, or to isomerism among the acids themselves.[1]

Barbier sought to find distinct chemical differences between the two and suggested that the minerals may be distinguished by the fact that orthoclase often contains traces of lithium and rubidium while microcline does not seem to carry them. Boeke[2] pointed out that the two salts may have a selective solubility for these rarer alkalies, or that these control the magnitude of the twinning. But such a distinction is not fundamental as is shown by Vernadsky[3] when he found an unquestioned specimen of microcline from the Ilmen Mountains that contained rubidium to the extent of 3.13 per cent of Rb_2O.

It is generally stated by those who believe that there is no physical-chemical difference between orthoclase and microcline that the specific gravities of the two are identical. It is true that they closely approximate each other, as would be expected in the case of minerals so closely allied, but the slight differences are significant when definitely charted after careful analysis as is shown by the diagram, Figure 7, on page 231. Let us defer final conclusions on this argument till we come to its more detailed examination.

Another difference between the two minerals is the varying values of the indices of refraction. Weinschenk has stated it very clearly when he calls attention to the fact that orthoclase has lower indices of refraction in all crystallographic directions than anorthoclase, while those of microcline are a little lower in one direction and a trifle higher in others.[4]

Still another difference indicating a fundamental distinction between orthoclase and microcline is the thermal constants of these minerals as shown in Table I.[5]

[1] F. W. Clarke, "Constitution of the Natural Silicates," *U.S. Geol. Surv. Bull. 588*, p. 35.

[2] H. E. Boeke, *Grundlagen der Physikalisch-Chemischen Petrographie* (1915), p. 158.

[3] Vernadsky, *Bull. Soc. Min.*, Vol. XXXVI (1914), p. 258.

[4] Weinschenk-Clark, *Petrographic Methods* (1912), p. 330.

[5] D. M. Liddell, *Metallurgists' and Chemists' Handbook*, second ed., 1918, pp. 207–8; Joseph W. Richards, *Metallurgical Calculations*, 1918, p. 140.

If the figures are reliable there should be some basis for the opinion that the two similar minerals may be isomeric forms of the same substance.

TABLE I

Mineral	Melting-Point	Latent Heat of Fusion	Specific Heat
Orthoclase..........	1200°	100 cals.	0.1877
Microcline..........	1170°	83 cals.	0.197*

*See W. P. White, *Amer. Jour. Sci.* (4), Vol. XLVII (January, 1919), p. 17, for more modern values of the specific heat of microcline.

Harker says:

> If orthoclase and microcline are dimorphous, the latter must clearly be the lower [temperature] form. Where it occurs with the apparent characters of a primary mineral it is the latest product of crystallization and is characteristic of the most acid of granites and especially of pegmatites. . . . . The conversion of orthoclase to microcline, or the setting up of microcline structure in orthoclase, has been attributed to dynamic causes.[1]

Another explanation of the microclinic texture in potash feldspar is that it is solely due to stresses set up by dynamic forces and is not explained by the theory of dimorphism. Thus Rosenbusch[2] remarks:

> The fact that microcline is almost wholly confined to the older eruptive rocks which have been subjected to processes of faulting and pressure, together with the observation that normal orthoclase assumes the microstructure of microcline when it has experienced strong pressure, leads to the supposition that microcline-structure is a pressure phenomenon. The correctness of this assumption can be verified in many cases. The occurrence of crystals of microcline in cavities, however, proves that it has not been produced in this way in all cases. Still the absence of microcline from unaltered extrusives is notable.

The writer is fully aware that microclinic texture is frequently the result of dynamic stresses, but is inclined to the view that the stresses of dynamic metamorphism permit orthoclase, which is metastable at temperatures below its inversion point, to change to microcline. His conclusion is that pressure does not produce microcline from orthoclase; it only initiates and accelerates the change.

[1] Alfred Harker, *Natural History of Igneous Rocks* (1909), p. 259.

[2] Rosenbusch-Iddings, *Microscopical Physiography of Rock-Forming Minerals*, p. 320.

Working with thin sections and crushed fragments of the same specimen it was frequently observed that while the thin section showed the feldspar frequently twinned the crushed fragments of the same material were often untwinned. This applies to plagioclase as well as to the potash-soda varieties. In fact some "adularias" and "microclines" when examined in the form of thin plates (removed from the specimen by a knife blade) do not exhibit twinning while in thin sections of the same specimen the characteristic twinning is at once manifest. An excellent example of this phenomenon is the "microcline" from San Diego County, California. The suggestion is strong that sufficient pressure to develop twinning was present in the grinding process necessary for the preparation of the thin section. If soda orthoclase is metastable at normal temperatures then it follows theoretically that inversion to soda microcline would be hastened by heating. This proved to be the case. The San Diego County material was crushed, sieved, and sized, and then divided into four samples. One sample was not heated. No. 2 was heated one hour in a quartz glass crucible over a special Bunsen (Scimatco) burner. No. 3 was heated three hours, and No. 4, five hours. It was found that the percentage of the twinned fragments increased with the duration of heating. The significance of this simple experiment is the raising of the question whether thin sections can be relied upon uniformly as a means of proper identification of the feldspars.

The experimental work of Allen and Day[1] shows that the determination of the melting temperatures of the alkali feldspars and the thermal-reaction points indicative of inversions is extremely difficult. Working with natural microcline from Mitchell County, North Carolina, and employing "thermal apparatus . . . . sufficiently sensitive to detect any unsteadiness of a tenth of a degree [centigrade] with certainty, not the slightest trace of an absorption or release of heat was found." All such "phenomena appeared to be effectively veiled by some property [of the substance], presumably the viscosity." In natural magmas the presence of mineralizers "acting as solvents, keeps the minerals in a fluid condition until the temperature is far below that at which they would

[1] Allen and Day, *Carnegie Institute Pub. 31*.

otherwise solidify, thereby making possible their crystalline development."[1]

While it seems as if the matter cannot be definitely settled by laboratory work unless fluxes or high pressures are employed, yet the writer is of the opinion that microcline is an isomer of orthoclase. This implies that above the transformation temperature orthoclase is the stable mineral, while below it microcline is the normal form. If orthoclase passes this point on cooling without inverting, then the mineral exists in a metastable condition.

THE SODIUM COMPONENT

Albite.—Albite (*albus*, Latin for white) is the soda feldspar, $NaAlSi_3O_8$, which crystallizes in the triclinic or anorthic system. Structurally it can be represented by the same general formulas given above by substituting Na for K.

Barbierite.—It was not until recently that a monoclinic form of the soda component was even suspected. Barbier and Prost[2] found a sodium feldspar that is monoclinic and isomorphous with orthoclase, which Clarke[3] lists as "barbierite isomeric with albite." Schaller[4] says that the existence of a monoclinic soda feldspar isomorphous with orthoclase must be admitted. Thus we may have a relationship in the soda component identical with that prevailing in the potash feldspar. The thermal ranges of barbierite are unknown.

THE CALCIUM COMPONENT

Anorthite.—Anorthite (ἀν, not, and ὄρθος, straight) is a triclinic lime feldspar, but unlike the other components it is usually regarded as a salt of orthosilicic acid (H_4SiO_4) instead of the trisilicic. The formula is commonly written: $CaAl_2Si_2O_8$ or $CaAl_2(SiO_4)_2$.

So far as known no isomeric form of the lime component exists, consequently the term "lime component" and "anorthite" can be used indiscriminately.

The above three components, potash, soda, and lime, are the most important feldspar constituents. Two other components however ought to be mentioned.

[1] Weinschenk-Johannsen, *Fundamental Principles of Petrology*, 1916, p. 41.

[2] "Sur l'existence d'un feldspath sodique monoclinique isomorphe de l'orthoclase," *Bull. Soc. Chem.* (1908), III, 894.

[3] F. W. Clarke, *U.S. Geol. Surv. Bull. 588*, p. 35.

[4] *Bull. Soc. Min.*, XXXIII (1910), 320.

THE BARIUM COMPONENT

Celsian.—Celsian $BaAl_2Si_2O_8$ "is monoclinic and isomorphous with orthoclase."[1] No triclinic isomeric form of celsian is definitely known.

CARNEGIEITE

Our knowledge of carnegieite, $Na_2Al_2Si_2O_8$, the polymer of nephelite, has been derived almost entirely from the researches of the Geophysical Laboratory of the Carnegie Institution, although Thugutt[2] obtained crystals of it by rapidly cooling "nephelite hydrate." Carnegieite was first recognized as occurring in a cinder cone on the island of Linosa,[3] east of Tunis. It is "chemically, as well as physically, a feldspar, differing somewhat from those belonging to the orthoclase-albite-anorthite series. It may represent a member of a new series, containing variable amounts of carnegieite."[4] It has the same chemical composition as nephelite but it differs therefrom by dimorphism. It is triclinic and often twins polysynthetically after the albite law and less frequently after the pericline law. Carnegieite is reported as having a specific gravity of 2.571 at 25° C. The reader is referred elsewhere for more details, especially to Bowen.[5]

No better summary of the feldspar components can be offered than by giving a modification of a tabulation suggested by Washington.[6]

Fieldspar Group

$(R'AlSi_3O_8)$ $(R''\ Al_2Si_2O)_8$	Monoclinic-triclinic

"Adular" Subgroup

$(R'AlSi_3O_8)$	Monoclinic
Orthoclase	$KAlSi_3O_8$
Barbierite	$NaAlSi_3O_8$
Celsian	$BaAl_2Si_2O_8$

"Albite" Subgroup

$(R'AlSi_3O_8)$	triclinic
Microcline	$KAlSi_3O_8$
Albite	$NaAlSi_3O_8$
Anorthoclase	$(K,Na)AlSi_3O_8$

"Anorthite" Subgroup

$(R''Al_2Si_2O_8)$	triclinic
Anorthite	$CaAl_2Si_2O_8$
Carnegieite	$Na_2Al_2Si_2O_8$

[1] F. W. Clarke, *U.S. Geol. Surv. Bull. 588*, p. 35, 1914.

[2] S. J. Thugutt, *Neues Jahrb.*, Beilage Band 9 (1894), p. 561.

[3] H. S. Washington, *Jour. Geol.*, XVI (1908), 10; H. S. Washington and F. E. Wright, *Amer. Jour. Sci.* (4), XXVI (1908), 187, and XXIX (1910), 52–70.

[4] J. P. Iddings, *Rock Minerals* (1911), p. 243.

[5] N. L. Bowen, *Amer. Jour. Sci.* (4), XXXIII (1912), 551–73.

[6] H. S. Washington, "Suggestion for Mineral Nomenclature," *Amer. Jour. Sci.* (4), XXXIII (February, 1912), 149.

The method of approach here is the application of the phase rule to the feldspar system. The original presentation of the phase rule was garbed in mathematical terms. For the involved equations as set forth by Gibbs the thermo-equilibrium diagram has been substituted. To the metallographer the construction and interpretation of these diagrams present no great difficulties, but to the average geologist they are conventions that possess little or no meaning. As an understanding of the diagrams of the feldspar system is essential to what follows, a section at the end of the paper in the Appendix is introduced in which their construction and interpretation are discussed.

For most purposes it is sufficient to consider that the binary systems, soda-lime and potash-barium feldspars, constitute a series of solid solutions, the thermo-equilibrium diagrams of which are classified by Roozeboom as Type I. On the other hand the binaries, orthoclase-albite and orthoclase-anorthite, are represented by the eutectiferous type of diagram. Thus the feldspar binaries may be classified as follows:

	Isomorphous Type (Type I)	Limited Solubility Eutectiferous Type (Type V)
System................	Soda-lime feldspars	Potash-soda feldspars
Name..................	Plagioclase series	Perthite series
System................	Potash-barium feldspars	Potash-lime feldspars
Name..................	Hyalophane series	No name*

*The writer proposes on a later page the term "oranite."

TWO-COMPONENT SYSTEMS

THE SODA-LIME FELDSPARS—PLAGIOCLASE SERIES

The soda-lime feldspars or the plagioclase series is the best known isomorphous series in the mineralogy of the rock-forming minerals. Tschermak in 1864 propounded the theory that the plagioclase feldspars are isomorphous mixtures of albite and anorthite as is indicated by the formula: $m(NaAlSi_3O_8)+n(CaAl_2Si_2O_8)$. Vogt developed this theory, and established it from indirect evidence. It was experimentally demonstrated by the classical work

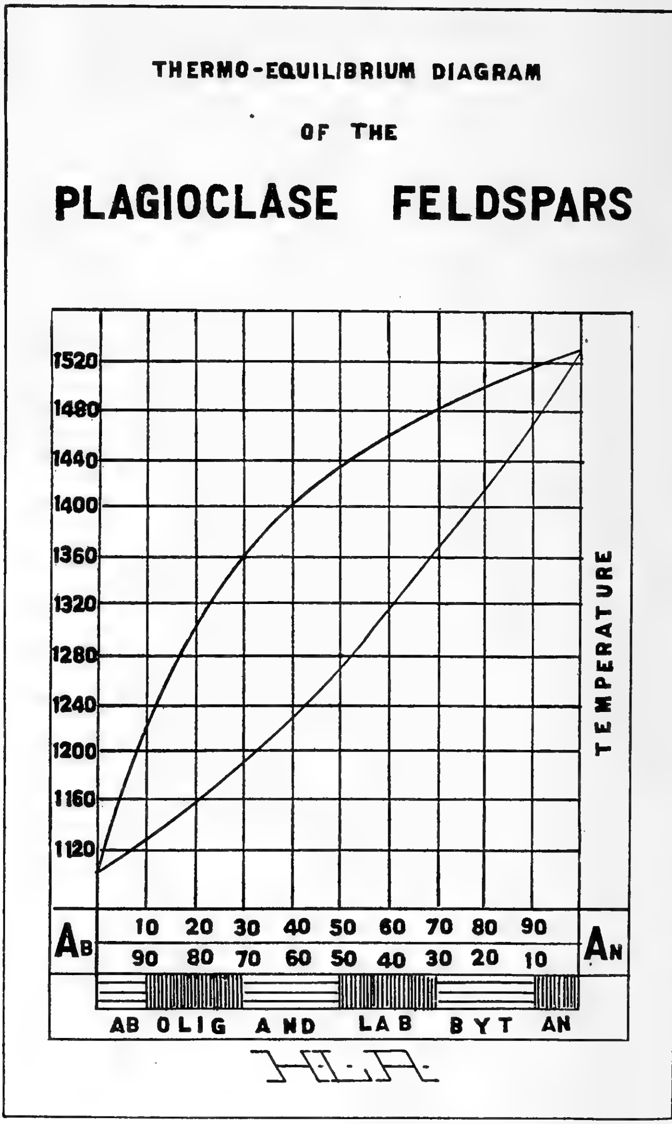


FIG. 2.—Thermo-equilibrium diagram of the plagioclase feldspars, based upon a molecular percentage (after Bowen).

of Allen and Day.[1] Figure 2 is the equilibrium diagram[2] of the soda-lime series, based upon a molecular percentage instead of a weight percentage as has been done in all the other diagrams.

The study of the solid-solution type of diagram shows us that the resulting crystals freezing from a mutual solution will not be homogeneous in composition, unless simultaneous or subsequent adjustment takes place, and consequently will vary in their optical properties as may be seen in zonal-grown crystals. But in deep-seated igneous rocks where the time of cooling is long, the non-homogeneous crystals gradually become uniform in composition by readjustment or exchange with each other and with the surrounding liquid. This process occurring between crystals is known as diffusion. All gradations between beautifully zoned plagioclase crystals and perfectly homogeneous ones occur in nature. The degree of homogeneity is therefore a function of the rate of chill, or, as the metallurgists would say, a measure of the rapidity of the quenching of the silicate alloy. Some zonal textures are, however, due to more complex processes such as the reabsorption of the margins of the already solid crystals and the adding of a new coating or layer deposited thereon. The writer believes that undue emphasis has been paid to the latter explanation of such textures and entertains the view that a large proportion of zonal crystals are not due to "magmatic corrosion," as it is called, but to normal magmatic crystallization under the influence of rapid chill.

Sometimes the crystal zones are sharply defined, each possessing fairly constant physical and chemical properties. These can be explained by irregularities in the rate of cooling.[3] The reversal of the order of zoning may be due to undercooling or the exposure of crystals to a liquid of a composition different from that of the one from which they crystallized.

The Properties of the Soda-Lime Feldspars.—It would seem as though our present knowledge of the physical properties of the plagioclase series was as complete as could be desired. Through

[1] A. L. Day and E. T. Allen, "The Isomorphism and Thermal Properties of the [Plagioclase] Feldspars," *Carnegie Inst. Pub. 31* (1905).

[2] N. L. Bowen, "Melting Phenomena of the Plagioclase Feldspars," *Amer. Jour. Sci.* (4), XXXV (1913), 583.

[3] N. L. Bowen, *Amer. Jour. Sci.* (4), XXXV (1913), 597.

the studies of Michel-Lèvy, Lacroix, Fouquè, Iddings, and many others, we have come to possess comprehensive data, more especially in regard to the optical properties. Under the circumstances it is

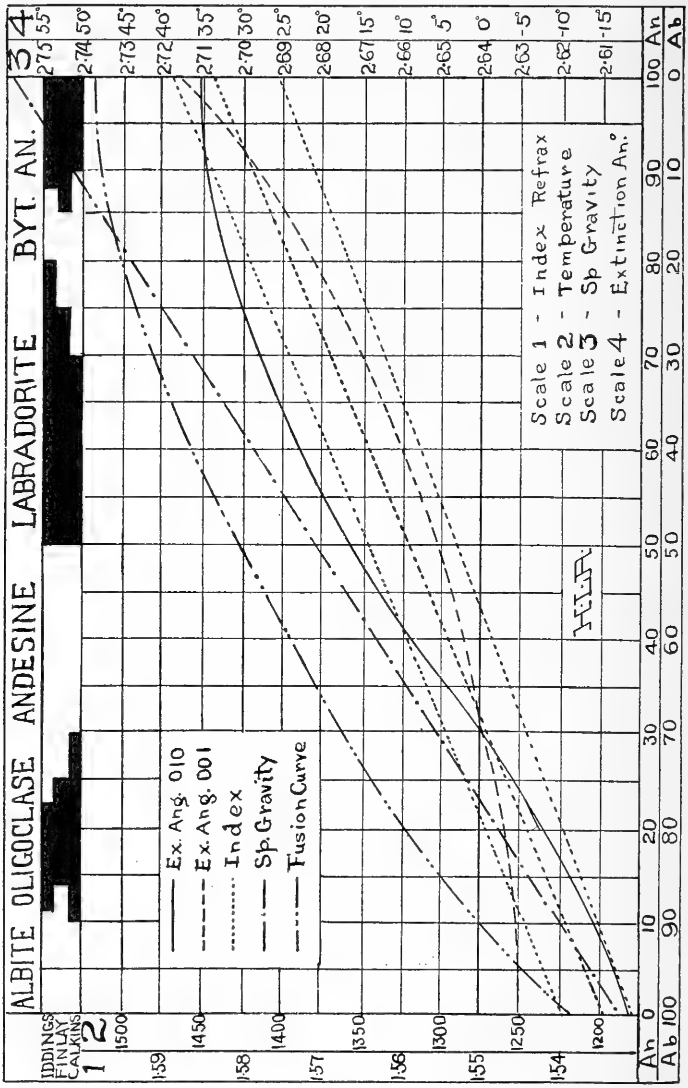


FIG. 3.—Plot of the physical properties of the plagioclase feldspars, together with the liquidus curve of the thermo-equilibrium diagram. The classification of the series is shown at the top of the figure.

rather a surprising fact that very few authors correlate or show the relation between these optical and physical properties and their less-known thermal properties.

Groth[1] says that the physical properties of a completely isomorphous series "are continuous functions of their composition." In fact it may be safe to regard a binary system to be isomorphous from a knowledge of these physical curves alone even though the system has not been subject to thermal investigation. Thus for the plagioclase series we can draw the following curves: (1) specific gravity, which in case of artificial feldspars is a straight line, (2) indices of refraction, alpha, beta, and gamma values, and (3) extinction angles on the (010) and (001) crystallographic faces. Although these functions have long been known to petrographers it may be that their presentation in graphic form, as shown in Figure 3, demonstrates relationships which hitherto have not been sufficiently emphasized.

Classification of the Plagioclase Series.—The plagioclase series is classified into six subdivisions, albite, oligoclase, andesine, labradorite, bytownite, and anorthite. These subdivisions are not a random group of minerals with definite composition arbitrarily classified into a group but constitute a series of steps or gradation from one end member of the series, albite, to the other end member, anorthite, and consist of these end members in reasonably definite but differing proportions. There is, however, a lack of agreement among petrographers as to the limits of the variations that may occur in these subdivisions without altering the mineral names which have been assigned to them respectively. F. C. Calkins[2] has lessened the difficulty by suggesting a decimal standard of composition as a basis of classification. This method, adopted by the writer, gives the following proportions as applied to the members of this series. It will be noticed that each of the minerals varies within definite limits as to the percentage of the two constituents which it may contain without losing its identifying name.

Albite,	from $Ab_{100}\ An_0$ to $Ab_{90}\ An_{10}$
Oligoclase,	from $Ab_{90}\ An_{10}$ to $Ab_{70}\ An_{30}$
Andesine,	from $Ab_{70}\ An_{30}$ to $Ab_{50}\ An_{50}$
Labradorite,	from $Ab_{50}\ An_{50}$ to $Ab_{30}\ An_{70}$
Bytownite,	from $Ab_{30}\ An_{70}$ to $Ab_{10}\ An_{90}$
Anorthite,	from $Ab_{10}\ An_{90}$ to $Ab_0\ An_{100}$

[1] Paul Groth, *Chemical Crystallography* (1906), p. 96.

[2] F. C. Calkins, *Jour. Geol.*, XXV, 157–59.

It will be observed that the term "albite" is used in two senses: as a component of the system, that is pure $NaAlSi_3O_8$; and as a mineral capable of a range in composition within definite limits. All mineralogists, consciously or unconsciously, use this dual nomenclature: (1) albite as a component; (2) albite as a mineral found in nature; a distinction necessary to make. No modification of the classification limits of the plagioclase series can reduce these two uses to one. The metallographer has a similar problem. His components are metals, his solid solutions and definite chemical compounds are "meterals."[1] Meterals are strictly analogous to minerals. Consequently we see that in mineralogy there is no term corresponding to the metallographer's term "metal." There would be a distinct gain if we possessed a name for the components of a mineralogical system. The writer is using the word "minal" to convey the meaning here expressed. Thus there is albite as a minal, and albite as a mineral.

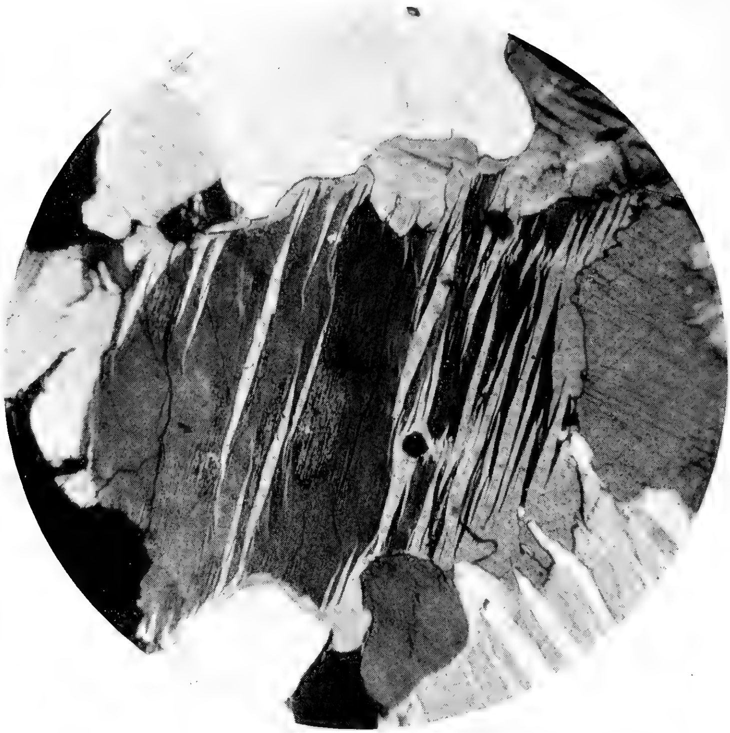
The theoretical percentages given above are based upon the assumption that the plagioclase series is a simple binary system. While such a conception is sufficient for most cases, a little study shows that natural specimens almost invariably contain some potash feldspar, the maximum being 10 to 12 per cent of the total in specimens near the albite end of the series, and decreasing as the percentage of anorthite increases. The potash component does not enter into the system in quite the same manner as the other two members for it is not a completely isomorphous component. Thus in classifying natural specimens one is compelled to consider the plagioclase series as a three-component rather than a two-component system.

Now the question arises what is the particular form of the potash feldspar which is found in the plagioclase feldspars? Miers[2] says

> since most of the plagioclase [feldspars] contain potash, we have to suppose either that monoclinic orthoclase can form [imperfect] isomorphous mixtures with triclinic plagioclase, or that the potassium feldspar is dimorphous, and that a modification exists belonging to the triclinic system and capable of entering into these mixtures.

[1] H. M. Howe, *Metallography of Steel and Cast Iron*, p. 232; Albert Sauveur, *Metallography and Heat Treatment of Iron and Steel*, pp. 293–94.

[2] H. A. Miers, *Mineralogy* (Macmillan, 1902), p. 452.



Microperthite (hyperperthite) in augite-syenite-granite, Ausable Forks, Clinton County, New York. Polarized light. ×50. Specimen 61.

The question is, is it orthoclase or microcline that occurs in plagioclase? We have learned from our discussion of isomorphism that it is probably microcline.

THE POTASH-SODA FELDSPARS—THE PERTHITE SERIES

Before the advent of Vogt's equilibrium diagram and Warren's masterful discussion of the potash-soda series it was held that this system was similar to if not identical with the plagioclase series; the two components being regarded as perfectly isomorphous.[1] But the Winchells[2] question this and say: "The close relation and gradation of optical properties corresponding to a gradation in chemical composition which exists in the plagioclase feldspars does not exist, or at least, has not been established, in the soda-potash feldspars." Harker[3] states that "we have to do with two . . . . interrupted series, Or-Ab, and Or-An, with a wide hiatus in the middle."

There would thus seem to be a hopeless difference of opinion regarding the potash-soda series. It will be shown later that these two extreme views are not as antagonistic as now appears.

Zapfee[4] is right when he says that "very little has been written regarding perthite, or perthitic intergrowth."

J. H. L. Vogt[5] was the first to give us an elaborate paper regarding the alkali feldspars, but his paper leaves many questions still obscure. Makinen[6] has studied the perthites from the pegmatites of Finland. But it has remained for Warren[7] to sum up the status of the perthite series and to discuss it quantitatively.

While the binary system, potash-soda feldspars, has not as yet been investigated thermally there seems but little doubt that

[1] P. Macnair, *Introduction to the Study of the Rocks and Guide to the Rock Collections in Kelvingrove Museum* (1911), p. 28; J. P. Iddings, "Obsidian Cliff, Yellowstone National Park," *U.S. Geol. Surv., Seventh Ann. Rept.*, pp. 269–70; A. H. Phillips, *Mineralogy* (1912), p. 408.

[2] N. H. and A. N. Winchell, *Elements of Optical Mineralogy*, p. 219.

[3] Alfred Harker, *Natural History of Igneous Rocks*, p. 244.

[4] Carl Zapfee, *Econ. Geol.*, VII, 137.

[5] J. H. L. Vogt, *Tschermak's mineral. und petrog. Mitt.* (1905), p. 24.

[6] E. Makinen, "Die Granitpegmatite von Tammela in Finnland," *Bull. d'l'Comm. Geol. de Finlande* (1913), pp. 1–101.

[7] Chas. H. Warren, *Proc. Amer. Acad. Arts. and Sci.*, LI (1915), 125–54.

it is of the general character shown in Figure 4 with the two components so regarded for the present only partially soluble in each other in the solid state, that it is a eutectiferous series. The first attempts to represent the diagram of the potash-soda feldspars were based upon the assumption that each component maintained uniform physical and chemical properties throughout heating and cooling. This assumption of course eliminated the possibility of their possessing dimorphous modifications. Consequently Vogt's diagram is extremely simple, and some of the lines are only approximate. Warren however suggested that the solubility lines should be drawn somewhat inclined instead of vertical as shown by Vogt. Harker illustrates by a diagram a eutectiferous system of which one component is dimorphous but did not apply it directly to the potash-soda feldspars. Warren stated the probability of the dimorphism of the potash component, orthoclase to microcline, and constructed his diagram with this in mind. Marc[1] has a different conception, which, although it possesses considerable merit, cannot be discussed in detail here. For the present we can derive considerable light upon the nature of the potash-soda feldspars by considering the diagram as given by Warren (Fig. 4).

In spite of the fact that the locations of the various points and the slopes of the curves of Warren's diagram are only approximate at the best it is believed that they are sufficiently accurate for our purpose. Warren has emphasized the truth that the two crystalline phases that make up perthitic intergrowths are solid solutions and not pure components. That is if we define orthoclase as pure $KAlSi_3O_8$ and albite as pure $NaAlSi_3O_8$ then the usual definition of perthite is not entirely satisfactory. Before the application of the phase rule to silicate systems the theory that the imbedded spindles had been introduced from without, subsequent to the solidification of the feldspar, could be regarded as plausible.[2] But today such an idea is abandoned by most workers in the field. In Figure 4 the composition of the original feldspar melt with an assumed value of 60 per cent K-feldspar and 40 per cent Na-feldspar is represented by the vertical line *NQC*. Now following

[1] Robert Marc, *Chemische Gleichgewichtslehre* (1911), p. 102.

[2] O. Wenglein, *Ing. Diss. Kiel*, 1903.

the method explained in the Appendix, the crystals will be found to have a composition of R through S toward A, as the temperature falls from T to T_2. The composition of the liquid on the other

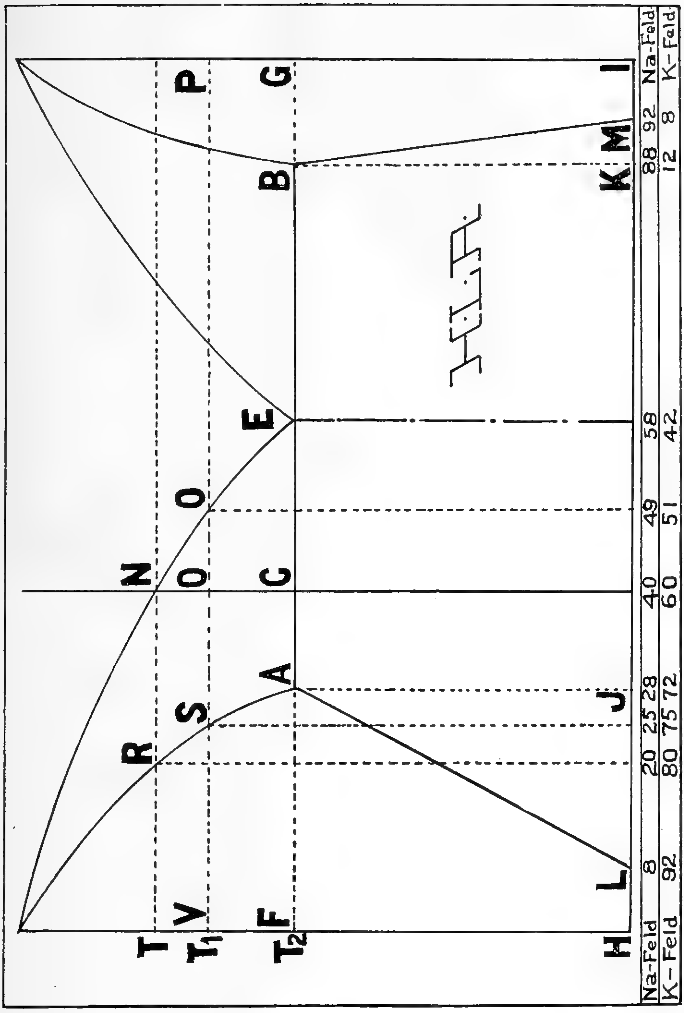


Fig. 4.—Thermo-equilibrium diagram of the potash-soda series, based upon a weight percentage (after Warren).

hand is indicated by the line NOE. Let us examine the crystal—a solid solution, which is of variable composition provided there is no readjustment taking place between the crystal phases and

the melt, each increment of which is in various degrees of saturation in respect to the sodium component. Then the phenomenon of diffusion (due to osmotic pressure) should take place tending to bring about a homogeneous mass. Let us suppose this diffusion takes place to completion while the temperature remains constant at T_2. A homogeneous solid solution nearly but not absolutely saturated with the sodium component results. We observe that the line AL is inclined, a fact which informs us that as the temperature falls the solubility likewise decreases. A temperature will soon be reached where saturation is complete. As "exsolution"[1] in solids is up to this point excessively slow, due to the high viscosity of that state, a condition of a supersaturation will occur. Assume that the feldspar is at normal temperature ($HLJKMI$), then in the course of geologic time this supersaturated crystalline mass will gradually separate into the two phases and give perthitic intergrowth. The perfect orientation of the crystal units in the host mineral would be maintained and hence we would expect to find that albitic phase "in irregularly lenticular layers . . . . in planes parallel to (801) or (100) and that both feldspars could have (010) in common."[2] This is probably the common form of perthite, due to exsolution of a supersaturated solid solution. As this is not a eutectic mixture it is well to be cautious in applying the term to all intergrowths found in rock sections. In fact many intergrowths that have the appearance of being eutectics may be due to secondary or subsequent processes. Whitehead[3] has emphasized this and illustrates the striking similarity in appearance between a true eutectic of 70 per cent of silver and 30 per cent of copper, and certain intergrowths which are found in sulphide

[1] There is no satisfactory word to express the phenomenon of the separation of two crystal phases due to supersaturation. "Precipitation" cannot be used as it designates a chemical relation; there is no chemical reaction taking place; it is merely a physical change. Warren uses the term "unmixing." This is not satisfactory for we are not dealing with original mixtures but with solutions. If we were observing a liquid solution we could say "crystallizes out of solution" but the homogeneous mass is already crystalline and such an expression would be unfortunate. It is really the opposite of "passing into solution," hence the term "exsolution" is proposed. The German term is "entmischung."

[2] J. P. Iddings, *Rock Minerals* (1911), p. 239.

[3] W. L. Whitehead, *Econ. Geol.*, XI, 1–13.

ores. His opinion is that the great majority of such intergrowths are due to secondary metasomatic replacement and not to the decrease in solubility of the components as they pass from the liquid to the solid phase. They are not eutectics in the sense in which the metallographer would use the term.

C. H. Smyth, Jr.,[1] says in regard to the microperthite contained in a "gneiss" from the Adirondack Mountains (which is, in all probability, the Adirondack augite-syenite of Algoman age):

> A very marked feature in a majority of the sections examined is the great abundance of the microperthite intergrowth of orthoclase and plagioclase. . . . . In most instances the microperthite has the appearance of that of a contemporaneous crystallization of the two feldspars; but enough sections contain absolute proof of its secondary nature to render it extremely probable that in this gneiss it is never an original intergrowth. Evidence of this secondary origin is seen in the plagioclase spindles passing unbroken across cracks in the orthoclase and in the evident optical continuity of the material of the spindles and secondary feldspar filling cavities and cracks adjacent to the microperthite.

Smyth gives the reader the impression that "contemporaneous crystallization of the two feldspars" and their "secondary origin" are incompatible. The writer would agree that the spindles could have formed subsequent to solidification and thus are secondary in this sense. Yet the material found in the two phases was present at the time the rock solidified. Smyth has apparently overlooked the question of the relative solubilities of the two phases with lowering temperature.

The examination of a large number of perthitic feldspars in thin section reveals the fact that most of them did not have an original eutectic composition, but have assumed their present form through the agency of exsolution. However, on the margins of the grains or in the interstitial spaces, perthite feldspars are frequently found, and are interpreted as representing the eutectic mixture. An excellent example of this type of feldspar is the pegmatite (graphic granite) from Bedford, New York.

Howe[2] makes the distinction between the eutectoid and the conglomerate of cementite (Fe_3C) and ferrite (alpha iron) in steels.

[1] C. H. Smyth, Jr., *Trans. New York Acad. Sci.*, XII (1893), 204.

[2] H. M. Howe, *Metallography of Steel and Cast Iron* (1916), pp. 71, 161.

The latter is the eutectic mixture. The eutectoid marks a change in the solid state and is called "pearlite," while the eutectic is referred to as "primary pearlite" or "ledeburite," a distinction well to make. The eutectoid pearlite is somewhat analogous to perthite due to exsolution. As this exsolution is a slow process the appearance of the two phases may not occur until many thousands of years after the freezing of the magma. There would be a gain for the sake of clearness if the literature of petrography can receive the term "perthoid" to refer to intergrowths of potash-soda feldspars due to exsolution, while "perthite" could be reserved for the true eutectic.

Undercooling.—Hitherto it has been assumed that each new phase made its appearance at the temperature at which it theoretically formed. In the laboratory this condition is rarely secured, while in nature, pressure, gases, and time are factors which profoundly affect the crystallization of a system. Inertia often causes a melt to remain in the liquid condition although the temperature may be below the freezing-point. This phenomenon is termed "undercooling."

The bridging of the eutectic gap by undercooling is discussed with the aid of the diagram shown in Figure 5. Three variables are represented: temperature, composition, and degree of equilibrium; and consequently a three-dimension model is necessary for representation. The back plane, $ABT_4\ ET_2$, is the normal diagram of the potash-soda feldspars, after Vogt and Warren. The plane nearest the reader is the metastable diagram when the two components A_1 and B_1 are completely soluble in the solid. The form assumed by the liquidus and solidus curves comes within Roozeboom's classification of Type III. The analogous points of the two diagrams are connected. The field of the rear plane where two solid phases are in equilibrium, *CDGOF*, becomes more and more restricted in passing to the front plane and comes to an end at E_1I. Stated in words, this indicates that the solubility of the two solid phases in each other becomes increasingly greater until complete solubility prevails. The converse of this is that if homogeneous crystals are formed upon freezing through the process of undercooling then because they are supersaturated in

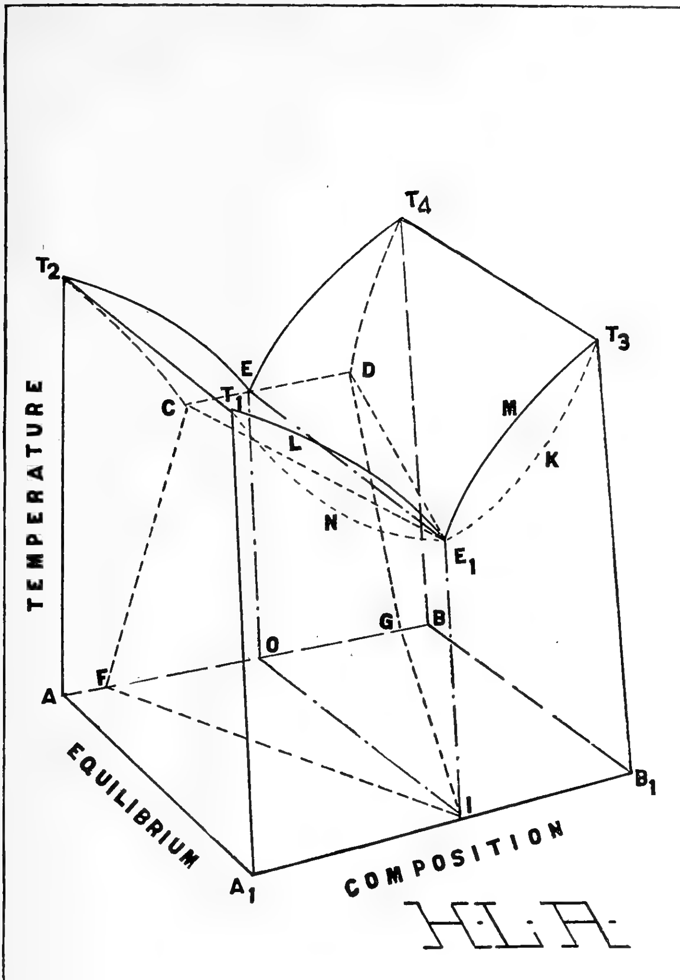


FIG. 5.—The stable thermo-equilibrium diagram of the potash-soda series after Warren, the back plane, the diagram for perthitic intergrowths and the metastable diagram, the front plane, for anorthoclase. The two diagrams are related through the degree of equilibrium.

respect to each other they will tend to separate into distinct and separate phases. The course of separation is indicated upon the basal plane of the diagram, by the lines *IF*, and *IG*, *F*, and *G* marking the limit. As a matter of fact we find in natural specimens every gradation between these two limits. That is, the diagram of one specimen may be properly represented by a cross-section of the solid model cut parallel with and near the back plane while another may be indicated by a section nearer to the front plane. It is obvious that the diagram, Figure 6, is highly diagrammatic and would be subject to modifications if the principle of dimorphism of the components were introduced.

Specific Gravities of the Potash-Soda Feldspars.—As has been noted before, Allen and Day have shown that the specific gravities of the plagioclase feldspars change uniformly from 2.605 for artificial albite, to 2.768 for artificial anorthite, the plot of their values assuming a straight line as in Figure 3. If, however, a break or cusp occurs in specific gravity curves it signifies a discontinuity in physical, crystallographic, and chemical properties, resulting from the development of new modifications or the formation of compounds. In the potash-soda series, the formation of compounds need not be considered, but the possible existence of dimorphous forms must be.

Hintze[1] gives the following data for the construction of such curves:

Mineral	Or-Ab Ratio	Percentage K_2O	Composition		Specific Gravity
			(A) Or Percentage	(B) K_2O Percentage	
Adularia.......	Or	16	100	95	2.56
Adularia.......	$Or_4\ Ab_1$	13	80	77	2.57
Amazonite.....	$Or_3\ Ab_2$	10	60	60	2.58
Perthite.......	$Or_1\ Ab_1$	7	50	41	2.60
Loxoclase......	$Or_1\ Ab_4$	4	20	24	2.61

In graphic form these data are ambiguous. The differences can be clarified by the examination of another system which has been studied experimentally: that of $FeSO_4 \cdot 7H_2O$ and $MgSO_4 \cdot 7H_2O$.

[1] Carl Hintze, *Handbuch die Mineralogie*, II, p. 1358.

Regers[1] found that the ferrous salt, when pure, is monoclinic in crystal habit while the magnesium salt is orthorhombic.

> From aqueous solutions, at ordinary temperature it is possible to obtain . . . . [solid solutions] containing up to 54 per cent of magnesium sulphate, and possessing the monoclinic form of ferrous sulphate. There must, therefore, be a second (a monoclinic) form of magnesium sulphate, which can form [limited] isomorphous mixtures with ferrous sulphate. There is a gap between the mixture containing 54 per cent of the magnesium salt and the next higher one, which contains 81 per cent, no intermediate mixtures being known. Mixtures from this point up to pure magnesium sulphate exist, and they exhibit the orthorhombic form of the latter salt. If the densities of these various mixtures are plotted against the corresponding percentage composition, as in Figure 6, it is seen that the values lie upon separate and distinct straight lines, not parallel to each other.[2]

The gap referred to above undoubtedly implies a limited solubility between the components; that is, limited isodimorphism. It is strongly suspected that the potash-soda feldspars are of this type. If so Hintze's data can be treated in the manner shown in Figure 6. The upper of these two non-parallel lines represents gravities of the monoclinic modifications, or orthoclase-barbierite; the lower line the triclinic forms, microcline-albite. If Hintze's data can be relied upon, the lines in the figure strongly suggest the dimorphism of each component, for it is known that a dimorphous substance possesses different gravities depending upon its modification, and that the density of the higher temperature form is almost always higher than that of the lower. Consequently we would expect to find that orthoclase has a higher specific gravity than microcline. This is apparently the case. A factor, however, that may nullify the latter interpretation, is that natural specimens of orthoclase almost invariably contain more sodium feldspar dissolved in them than is contained in microcline. Since the gravity of the sodium component is higher than either orthoclase or microcline this would increase the gravity of natural specimens of orthoclase above that possessed by microcline irrespective of any dimorphism that may exist. Until laboratory experiments are undertaken final conclusions are impossible.

[1] Regers, Various articles in *Zeitschr. für Chem.*, 1889, 1890, and *Jour. C.S.*, 1891.

[2] Paul Groth, English translation of *Einleitung in die chemische Krystallographie*, *Chemical Crystallography* (1906), p. 92.

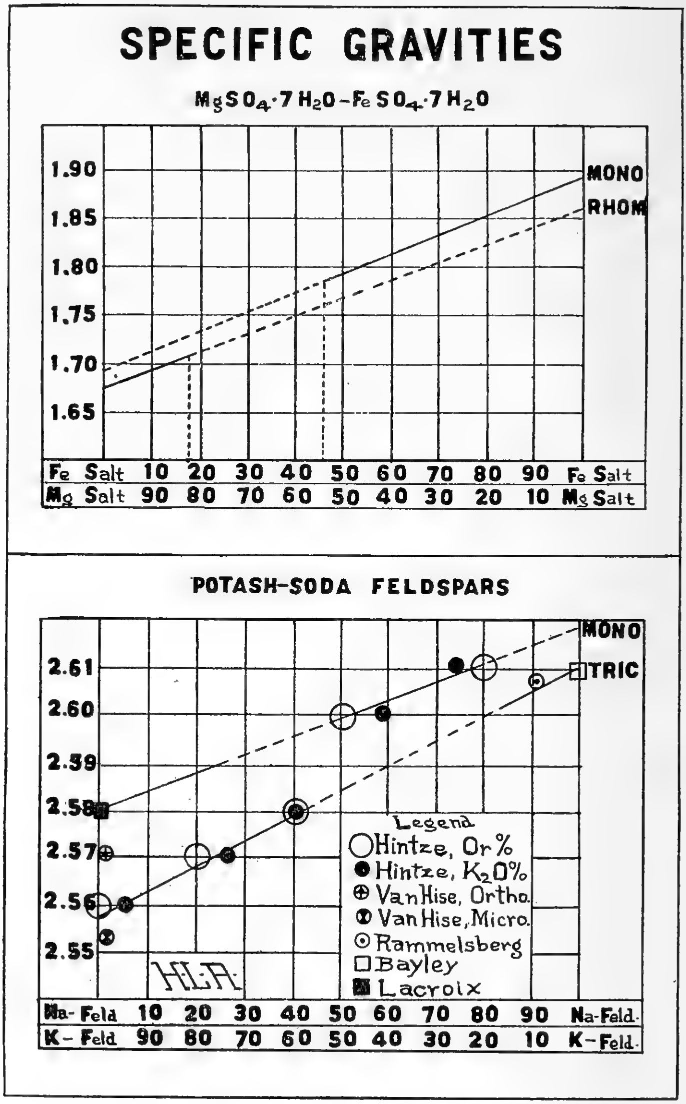


FIG. 6.—The specific gravities of the hydrated sulphates of iron and magnesium after Regers, and the specific gravities of the potash-soda series based upon data supplied by Hintze. Both diagrams illustrate the fact that the densities of dimorphous forms are different.

The Optical Properties of the Potash-Soda Feldspars.—It is a surprise, even to an experienced petrographer, that no adequate tabulation of the optical properties of the potash-soda series exists. What information there is in the literature is fragmentary and unrelated to the actual chemical composition. The writer has attempted to synthesize the data so given and has carried out sufficient microscopic examination of thin sections and crushed fragments to lead him to believe that it is possible to identify the subspecies of the potash-soda series with an accuracy approaching that obtained in the plagioclase feldspars.

Although the ease of distinguishing and identifying the different ranges in the plagioclase series is not duplicated in the potash-soda series, yet the maximum extinction angles on the (010) and (001) faces appear to offer fairly reliable criteria. In most textbooks on optical mineralogy the extinction angles for orthoclase on the (001) and on the (010) faces are given as 0°, and 4°30′ to 5°, respectively. As most, if not all, natural orthoclase contains some sodium feldspar in solution, the angular value for the pure substance is perhaps a little lower than the last named. As the amount of the sodium component increases to about 25 per cent the extinction angles on the (010) face increase from 5° to about 12° while the (001) extinction angle changes from zero to about 3°. In the form of crushed fragments this test becomes a simple though not a quick method of determination. A similar relation exists for microcline and soda microcline. The extinction angle on the (001) face is slightly less than 15° for nearly pure material but becomes 17° to 18° with increased soda content; likewise the (010) face extinction increases from 4°30′ to nearly 10°.

Potash albite is usually distinguished by the fine albite twinning, although this may be replaced by fine microclinic twinning instead, or no twinning may appear at all even under high magnifications, and reliance upon the extinction angles from cleavage faces must be substituted. The extinctions on the (001) face are about 5° and on the (010) face about 10°. With increasing soda content pure albite is reached when 21° on the (010) is found. The great difficulty is in distinguishing by extinctions alone saturated (or supersaturated) soda orthoclase from potash albite

containing its maximum amount of the potash component, but recourse to indices of refraction furnishes a fairly accurate means. The optical mineralogy of this system should be studied in conjunction with the chemical analyses, which for convenience should be recast in terms of the three components.

It is chiefly from the examination of some sixty specimens of natural potash-soda feldspars that the writer offers the diagrams (shown in Fig. 7) of the extinction angles of the potash-soda series. They are tentative and suggestive, nothing more. The dimorphism of the sodium component is not represented; for no information is at hand. Consequently barbierite has been left out of consideration.

Some of the uncertainty about the accuracy of the curves is because of the degree to which the lime component affects the extinctions. All of the specimens examined contained anorthite in solution up to 10 or 15 per cent of the whole and the problem was to correct the extinctions for the simple binary. The extinction angles of the plagioclase series, together with those of the potash-soda feldspars, whose chemical composition was known,[1] enabled the writer to plot equal extinction contours (isogonic lines) for that portion of the ternary system (K-, Na-, and Ca-feldspars) represented by natural specimens. These in turn made it possible to draw the extinction curves for the pure binary system.

Classification of the Potash-Soda Series.—Most mineralogists recognize that orthoclase usually contains an appreciable amount of soda, and that albite carries some potash. Winchell,[2] for example, gives orthoclase, $KAlSi_3O_8$; soda orthoclase, $(K,Na)AlSi_3O_8$; albite, $NaAlSi_3O_8$; and anorthoclase, $(Na,K)AlSi_3O_8$. While these expressions in general convey fairly definite meanings, it is well to remember that there is no definite ratio between the K and Na in homogeneous crystals which are members of the series, and that no single chemical formula can be given for them. Such feldspars with the potash and soda components in appreciable amounts should be considered as solid solutions limited to certain

[1] Dana, *System of Mineralogy*, sixth ed. (1892), p. 324.

[2] N. H. and A. N. Winchell, *Optical Mineralogy* (1909), p. 219.

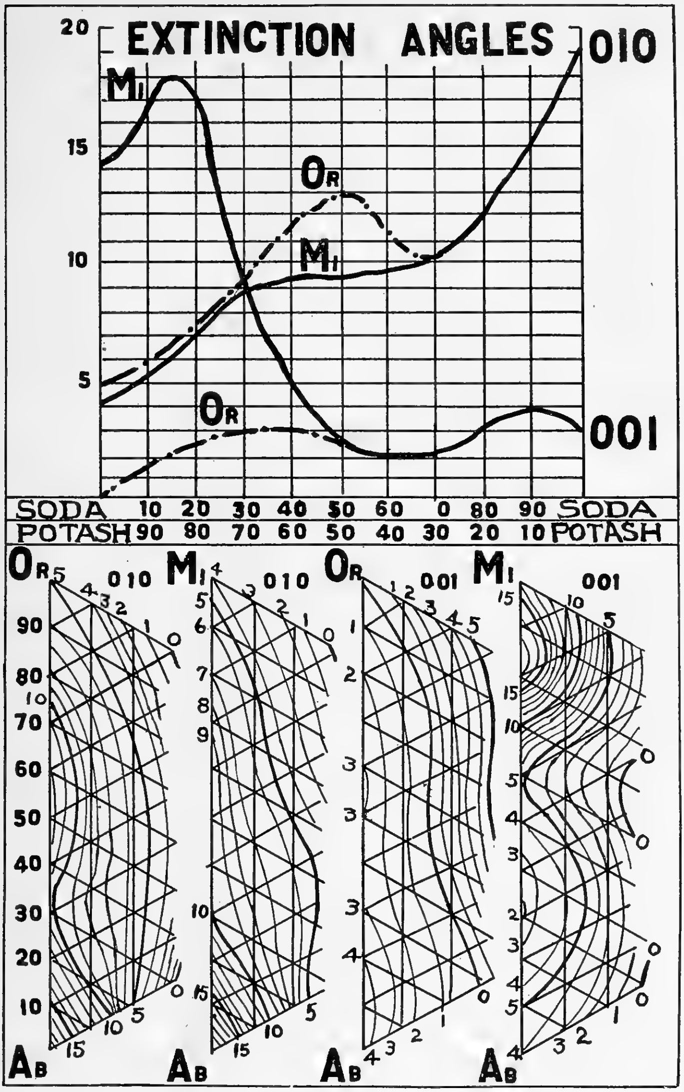


FIG. 7.—Extinction angles of the potash-soda series. The upper diagram shows values of the extinctions for the simple binaries, microcline, Mi; albite, Ab; and orthoclase, Or; albite, Ab, on the two faces, (010) and (001). The lower diagram shows the alkalic portions of the ternary diagram, K-, Na-, Ca-feldspars. The values of the extinction angles are shown by isogenic lines. Both diagrams are tentative.

ranges of composition. In general the term "soda orthoclase" is clear in its meaning.

Rogers[1] points out a danger in employing the term soda orthoclase in that it is somewhat ambiguous, for "soda-orthoclase may mean an orthoclase in which a portion of the potassium is replaced[2] by sodium or it may mean that sodium compound corresponding to orthoclase." For the latter it is better "to use a distinctive name for the monoclinic feldspar in which sodium predominates molecularly[3] over potassium." For such a mineral the term barbierite has been proposed. The writer fully appreciates the necessity for such distinctions, and maintains that soda orthoclase should mean an orthoclasic feldspar with some sodium component dissolved in it (Winchell's $[K,Na]AlSi_3O_8$) while sodium orthoclase should refer to barbierite.

The limits of the range here proposed for soda orthoclase are K-feldspar 90, Na-feldspar 10, to K-feldspar 70, Na-feldspar 30.

There exists considerable uncertainty regarding the composition of anorthoclase. Rosenbusch[4] gives as the range of this mineral the following ratios: Na-feldspar 67, K-feldspar 33-Na-feldspar 82, K-feldspar 18. These ratios suggest that anorthoclase is analogous to soda orthoclase, that is, it is a feldspar consisting chiefly of the sodium component with an appreciable amount of potash feldspar dissolved in it. Yet a study of the available chemical analyses of "anorthoclase" would suggest that these limits should be extended farther toward the potash side of the diagram, embracing in many cases the range where, under equilibrium conditions, perthite occurs. It is quite reasonable therefore to consider that many anorthoclases are undercooled metastable solid solutions of the two alkali components. A clear

[1] A. F. Rogers, "The Nomenclature of Minerals," *Proc. Amer. Phil. Soc.*, LII (1913), 610.

[2] It is well to recall the objection to an expression of this kind.

"Replaced" is unsatisfactory in a physical-chemical sense.

Professor Johannsen informs me that he has used "proxied by" instead of "replaced by." There is no misunderstanding of that term.

[3] The application of X-rays to crystal structure has shown that molecules do not exist as such in solids.

[4] Rosenbusch-Iddings, *Microscopic Physiography of the Rock-Forming Minerals*, p. 340.

conception, such as this, seems necessary to supplant the present nomenclature which is very unsatisfactory due to its indefiniteness.

The question arises whether it is best to restrict "anorthoclase" to the central portion of the equilibrium diagram and to undercooled homogeneous crystals of perthitic composition, or to limit it to albitic feldspars containing the potash component up to a maximum of 20 per cent. After due consideration the former proposal is the one here adopted. The average of the 47 available analyses of this mineral is higher than 20 per cent in the potash component and is therefore within the perthitic range.

If the term "soda orthoclase" is used for that portion of the potash-soda series indicated by the range $Or_{90}Ab_{10}$-$Or_{70}Ab_{30}$ and the term "anorthoclase" for the range in the center of the equilibrium diagram, then it is necessary to supply a term for albitic feldspars containing the potash component up to 20 per cent. For this range of solid solutions, K-feldspar 20, Na-feldspar 80- K-feldspar 5, Na-feldspar 95, the term "potash albite" is proposed.

Anorthoclase is here used as the name to designate the range, K-feldspar 70, Na-feldspar 30- K-feldspar 20, Na-feldspar 80, when it is a supersaturated undercooled metastable homogeneous solid solution, potentially perthite, through the intermediate stages of cryptoperthite and microperthite.

The emphasis that mineralogists place upon the distinction between monoclinic and triclinic crystals has resulted in considerable confusion about the distinction between soda orthoclase, anorthoclase, and soda microcline. For example, the interesting classification of Klockmann[1] will illustrate this point:

Formula	Monoclinic		Triclinic	
	Component*	Isomorphous Mixture	Component*	Isomorphous Mixture
$KAlSi_3O_8$....	Orthoclase	Soda Orthoclase	Microcline	Anorthoclase
$NaAlSi_3O_8$...	Unknown†		Albite	
$CaAl_2Si_2O_8$...	Unknown	....	Anorthite	Plagioclase

*Free translation of "Selbständig."

† For this modification of the sodium component the term barbierite has been proposed.

[1] Klockmann, *Lehrbuch der Mineralogie* (1912), p. 488.

He furthermore puts in parentheses after "anorthoclase," "soda microcline"[1] and "microcline albite."[2] A similar procedure is followed by Iddings[3] who speaks of the potash-soda feldspars rich in soda as "soda microcline (anorthoclase)."

It seems to the writer that this confusion is due to the failure to recognize the possible dimorphism of each of the components in the series, that is, there may be two distinct binary systems in one: the so-called monoclinic series with orthoclase and barbierite as end members, and the triclinic potash-soda feldspars with microcline and albite as components. If this is recognized then it is logical that the two should be classified as follows:

Monoclinic: Orthoclase, soda orthoclase, monoclinic anorthoclase, potash barbierite, and barbierite.

Triclinic: Microcline, soda microcline, triclinic anorthoclase, potash albite, and albite. The interrelationship that exists between the monoclinic and triclinic anorthoclases has been pointed out by Dana[4] who says that the axial angle of anorthoclase varies with the temperature, "becoming monoclinic in optical symmetry between 86° and 264° C. but again triclinic on cooling. This is true of those containing a little calcium" (anorthite). Although the writer is somewhat cautious in proposing new terms—for the literature of mineralogy and petrography is already burdened with many useless names, some of which are worse than useless—yet there has been no systematic attempt to subdivide the potash-soda series into definite ranges analogous to oligoclase, andesine, labradorite, etc. Calkins' decimal principle as applied to the plagioclase series is so logical that its application to the potash-soda series is worth attempting.

The terms hypoperthite and hyperperthite may appear strange to the petrographer but the metallographer will recognize old friends. The nomenclature proposed is an adaptation of the terms hypoeutectoid and hypereutectoid as applied to steels. The word eutectoperthite is self-explanatory. If such refinement in classifying the system is neither possible nor desirable, then perthite can

[1] Natronmikroklin. [2] Mikroklinalbite.

[3] Joseph P. Iddings, *Rock Minerals*, 1911, p. 235.

[4] James D. Dana, *System of Mineralogy*, sixth ed., 1892, p. 324.

be substituted for the three names in the center of the table as is indicated at the sides.

Monoclinic		Ratios	Triclinic	
	Orthoclase	$Or_{100}Ab_{0}$ -$Or_{90}Ab_{10}$	Microcline	
	Soda orthoclase	$Or_{90}\ Ab_{10}$ -$Or_{70}Ab_{30}$	Soda microcline	
Perthite (stable) Anorthoclase (metastable)	Hypoperthite	$Or_{70}\ Ab_{30}$ -$Or_{45}Ab_{55}$	Hypo-perthite	Perthite (stable) Anorthoclase (metastable)
	Eutectoperthite	$Or_{45}\ Ab_{55}$ -$Or_{35}Ab_{65}$	Eutecto-perthite	
	Hyperthite	$Or_{35}\ Ab_{65}$ -$Or_{20}Ab_{80}$	Hyperthite	
	Potash barbierite	$Or_{20}\ Ab_{80}$ -$Or_{5}\ Ab_{95}$	Potash albite	
	Barbierite	$Or_{5}\ Ab_{95}$ -$Or_{0}\ Ab_{100}$	Albite	

Some criticism may arise in that the above classification is more detailed than is warranted by the determinations possible with the petrographic microscope upon natural specimens. The writer feels, however, that with greater care and proper emphasis, microscopic distinctions are possible that will approach the accuracy now obtainable in classifying the soda-lime series.

THE POTASH-LIME FELDSPARS—"ORANITE" SERIES

One of the strange facts of mineralogy is that the potash-lime feldspars are considered rare in nature, and are little discussed in the literature. Harker[1] says "the relation between orthoclase and anorthite are doubtless of the same general kind [orthoclase-albite], though the higher melting-point of the latter mineral will presumably throw the eutectic point somewhat nearer to orthoclase." Bayley[2] remarks that "mixtures of the potash and calcium molecules[3] are extremely rare as minerals, but that they have been formed experimentally in the laboratory." All available data lead us to conclude that the $KAlSi_3O_8$-$CaAl_2Si_2O_8$ system is similar to the potash-soda series, except that we do not have to consider an isomeric modification of anorthite, which reduces the system to one of less complexity. The eutectic temperature is probably higher and the lines of solubilities are nearer to the sides of the diagram.

[1] Alfred Harker, *Natural History of Igneous Rocks*, p. 246.

[2] Wm. S. Bayley, *Descriptive Mineralogy*, 1917, p. 408.

[3] "Components" is to be preferred for reasons already given.

The reason why it is rare in nature is not far to seek; a rock high in potash and lime but low in soda is rare, and when found the two feldspathic phases occur as separate and distinct grains. This can be explained on the basis that melts of the potash-lime feldspars, especially when rich in the latter component, are much less viscous than the potash-soda ranges. Consequently their separation into distinct identities is much more common, and intergrowths are rare. That the latter do occur, however, the writer feels convinced.

In the Adirondack Mountains the anorthosite and the augite syenite, both of Algoman age,[1] occur in igneous contact with each other with such field relations as to indicate that certain batholithic masses of these two rocks invaded the country rocks (chiefly Grenville sediments) at about the same time, although the anorthosite is probably the older of the two in all cases. Magmatic assimilation of the syenite by the anorthosite has taken place to some extent in zones where they adjoin one another. The feldspar of the syenite is microperthite (antiperthite) (hyperperthite) while that of the anorthosite is acid labradorite ($Ab_{55}An_{40}Or_5$). In the syntectic rock, marking the zone of contact, microscopic examination reveals intergrowths of potassic feldspar and labradorite or orthoclase with an appreciable amount of lime ("lime-orthoclase") holding blebs of either labradorite or bytownite. It is obvious that such rocks are not "typical" and that the process of the development of these potash-lime feldspars is more involved than the simple freezing of a magma and consequently these cannot be pointed to as good examples of this rather neglected family.

But to return to more normal rocks, we find that the feldspathic content of most granites, syenites, monzonites, etc., is not limited to one species of feldspar. Even a hasty petrographic study of slides of these rocks shows "orthoclase and plagioclase." In the monzonites and granodiorites both alkali and plagioclase feldspars are present. Occasionally a basic representative of these carries in addition to the potassic or alkali feldspars, basic plagioclase,

[1] H. L. Alling, "Some Problems of the Adirondack Pre-Cambrian," *Amer. Jour. Sci.* (4), XLVIII (July, 1919), 62; "Geology of the Lake Clear Region," *N.Y. State Mus. Bull. 207–8* (1919), pp. 119–20.

labradorite, or even bytownite. In such rocks the feldspars are undoubtedly approaching, as a limit, the potash-lime binary system. The infrequency of intergrowths of these feldspars is the cause of the failure to recognize the system in nature.

Referring to the triangular plot of the feldspar analyses (see Fig. 19), it will be seen that a few specimens called "labradorite" and "anorthite" are approaching the side of the triangle occupied by the potash-lime feldspars.

The lack of specimens of this binary system prevents any accurate attempts being made to outline or to plot the physical and optical properties. They may, however, be inferred. The specific-gravity curves are steeply inclined, extending from 2.58 for orthoclase to 2.765 for anorthite. The gravity line for microcline and soda-microcline probably does not reach the high lime ranges, and ends somewhere between the two limits. The indices of refraction likewise are more inclined. In the center of the diagram, these lines have no practical significance, as undercooled metastable crystals of analogous to anorthoclase are probably unknown or very rare in normal rocks.

When we come to the nomenclature and classification of the binary, we enter virgin fields. What name shall be applied to the intergrowths described above as occurring in the Adirondacks? Shall perthite be used? There is considerable objection to such a practice; it would be extending the meaning of a well-established term and the modern tendency is in the opposite direction. Rogers[1] has proposed a slight extension of the term perthite to include intergrowths of two alkali feldspars when the orientation of the blebs differs from the customary relation. This implies that a certain element of textural habit is associated with the compositional significance. All of this leads to the obvious conclusion that a new word is required. The writer proposes, for intergrowths of potassic and high lime plagioclase feldspars, either eutectics or due to exsolution, the term "oranite." Its derivation tells the composition: *or*thoclase-*an*orthite—*ite*, the mineralogical ending. The same name can include intergrowths of soda microcline and potash anorthite.

[1] A. F. Rogers, *Jour. Geol.*, XXI (1913), pp. 202–7.

It will be seen that in order to limit the feldspar species to definite areas in a three-component ternary diagram the potash-lime series must be classified, in analogous fashion, to that already proposed for the soda-lime and potash-soda binaries. For feldspars of the K-Ca series, analogous to soda orthoclase and potash albite, the terms "lime orthoclase" and "potash anorthite," respectively, are suggested as meeting the present requirements.

THE BARIUM-POTASH FELDSPARS—HYALOPHANE SERIES

The name hyalophane is said to have been proposed by Waltershausen in 1855[1] from the Greek, ὕαλος, "glass," and φαινεσθαι, "to appear," alluding to its transparency. It is described as a barium-bearing feldspar found in transparent crystals similar to adularia. It is exceedingly rare in nature, the most famous locality being in the white dolomite of Binnenthal, where it was formed as the result of igneous contact action.

The melting-point of celsian is not known, and furthermore very few suggestions have been found in the literature that would indicate the probable nature of the thermo-equilibrium diagram. Iddings gives: $m(KAlSi_3O_8)$, $n(BaAl_2Si_2O_8)$, suggesting isomorphism similar to that possessed by the plagioclase series. Winchell gives a plot of the indices of refraction which incline toward higher values with increasing barium content, which is strictly analogous to the plagioclase system. Iddings supplies data for the construction of the specific-gravity curve, which takes the form of a straight line. Clarke[2] quotes the opinion of Standmark[3] that "the mineral celsian . . . . is monoclinic and isomorphous with orthoclase." Klockmann[4] likewise expresses the same view. In speaking of hyalophane in a more restricted sense, Clarke says: "Hyalophane and other barium feldspars are mixtures of orthoclase and celsian." In consistence with this view both Winchell and Iddings employ symbols and ratios which show a continuous range from Or to Cn, as follows: $Or_{19}\,Cn_1$, $Or_{10}Cn_1$, Or_4Cn_1, Or_3Cn_1, Or_7Cn_3, and Cn (Fig. 8).

[1] W. S. Waltershausen, *Pogg. Ann.*, XCIV, 134.

[2] F. W. Clarke, *U.S. Geol. Surv. Bull. 588* (1914), p. 35.

[3] Standmark, *Zeitschr. für Kryst. und Min.*, Vol. XL (1907), p. 89.

[4] Klockmann, "Isomorphe mischung von $K_2Al_2Si_6O_{16}$ mit $BaAl_2Si_2O_8$," *Lehrbuch der Mineralogie* (1912).

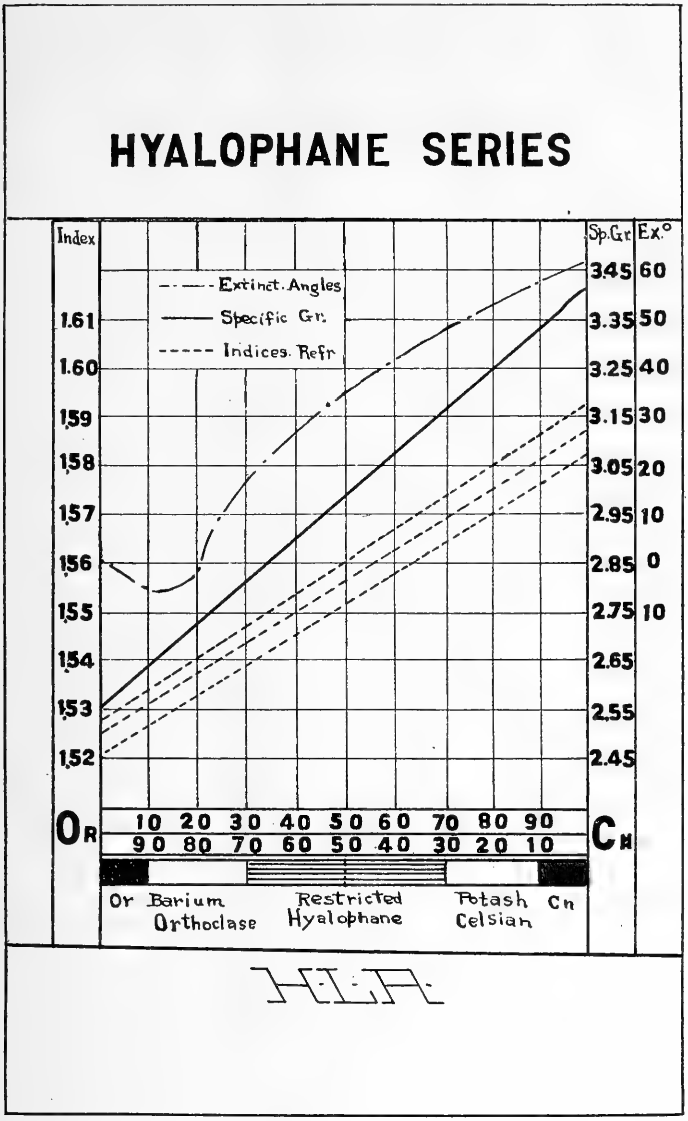


FIG. 8.—Plot of the physical properties of the hyalophane series. The uniform character of the curves implies that the orthoclase celsian system is a series of solid solutions, and that no chemical compound exists between the components. Data taken from Iddings and Winchell.

On the other hand the conception of a definite chemical compound within the series has been expressed by a number of mineralogists. Miers[1] says

> hyalophane corresponds to the formula $K_2BaAl_4Si_8O_{24}$ (not isomorphous $[K_2,Ba]$) and this can be expressed as a mixture of $BaAl_2Si_2O_8$ with 2 $(KAlSi_3O_8)$, i.e., as compounded of two molecules of orthoclase with one molecule of barium silicate similar to anorthite. This union is exactly analogous to the mixture of albite and anorthite in the (plagioclase) group . . . . but in hyalophane the mixture appears to be only in one definite proportion, so that the mineral is to be regarded as a double salt rather than a solid solution.

A similar view is taken by Moses and Parsons[2] who give hyalophane the following formula: $(K_2,Ba)Al_2(SiO_3)_4$ as though the fundamental acid was metasilicic. Let us see how such an interpretation is possible. If we employ one unit of the barium and two of the potash components then

$$2\times(KAlSi_3O_8)=K_2Al_2Si_6O_{16}, \text{ a trisilicate}$$

$$\frac{BaAl_2Si_2O_8}{K_2BaAl_4Si_8O_{24}}, \text{ an orthosilicate.}$$

Now

$$\frac{K_2BaAl_4Si_8O_{24}}{8}=K_2BaAl_4(SiO_3)_8.$$

And furthermore

$$\frac{K_2BaAl_4(SiO_3)_8}{2}=(K_2Ba)Al_2(SiO_3)_4,$$

which looks like a metasilicate but it may be far from being one. By taking two trisilicate units and one orthosilicate unit we get the result. The writer must take exception to such an interpretation. Furthermore Moses and Parsons indicate by the comma between the K_2 and Ba that the ratio between them is not constant. In other words the ratio of the number of units of the two components varies. If it varies then the so-called formula would be much more complex and depart from the form assumed by a metasilicate.

[1] H. A. Miers, *Mineralogy*, p. 461.

[2] A. J. Moses and C. L. Parsons, *Mineralogy, Crystallography, and Blowpipe Analysis*, fifth ed., p. 493, 1916.

Their formula, upon the basis that the end members are orthoclase and celsian, is misleading.

Groth[1] discusses the difficulty of distinguishing between isomorphous "mixtures" and compounds when there appears to be a definite and fixed ratio between the end members of a given series. In referring to the "triclinic feldspars" [plagioclase?] he says:

> The predilection toward certain definite mixture ratios in the series named is probably connected with the fact that apparently a regular distribution of the two kinds of atomic groups provides a particularly stable equilibrium of the crystal structure, since it occurs also with isomorphous substances of completely analogous chemical constitution.[2]

The fact that many "isomorphous mixtures [occur] in simple stoichiometric proportions appear in certain cases to possess greater stability than do those in other proportions." The modern "view of crystal structure . . . . shows . . . . that the formation of a crystal from two different kinds of chemical molecules,[3] even though these differ very slightly from each other, will give a particularly stable structure when the molecules[3] take part in this formation in regularly alternating manner; since such a substance has as much right to the name of 'molecular compound' as to that of 'isomorphous mixture,' it is evident that that view does not permit of any sharp boundary between the two ideas."[4]

Groth points out the inherent failure of the usual mineralogical methods of attack to distinguish compounds from isomorphous mixtures, solid solutions. Upon such problems mineralography sheds considerable light. Unless a compound is unstable at its melting temperature, the *liquidus* curve assumes a maximum in the form of an arch. In all cases there is an abrupt change in the direction and slope of the curves showing the physical properties of the series. This is clearly emphasized by the $CaSiO_3$-$MgSiO_3$ (Pseudowollastonite-Clinoenstatite) diagram.[5] Both of the end members of the series form eutectiferous mixtures with the common double salt, diopside. If the specific-gravity curve is plotted

[1] P. Groth, *Chemical Crystallography*, translation by H. Marshall, 1906, p. 98.

[2] P. Groth, *Neues Jahrb. f. Mineral.*, II (1903), 93 ff.

[3] Components. [4] P. Groth, *Chemical Crystallography* (1906), p. 105.

[5] Allen and White, *Amer. Jour. Sci.* (4), Vol. XXVII (1909); Ferguson and Merwin, *Amer. Jour. Sci.* (4), Vol. XXXVIII, August, 1919.

in conjunction with the diagram, it will be observed that it experiences a sharp change in direction at the point indicating diopside.

Now in the hyalophane series, we possess sufficient data to show that the specific-gravity curve is straight and therefore that there is no "molecular compound" between orthoclase and celsian. Consequently we must reject the theory stated by Miers, Moses, and Parsons as untenable.

THREE-COMPONENT SYSTEMS

TERNARY DIAGRAMS

In dealing with two-component systems we have to consider two variables—the temperature and the composition. This necessitates the employment of two-dimensional diagrams. But with three components we must employ three dimensions, in other words, solid figures. These are bulky affairs which are difficult to represent upon paper. There are two methods by means of which this may be accomplished: first, by perspective drawings (special form of "block diagram"); or second, by plan drawings, ignoring the vertical co-ordinate (temperature), and projecting the liquidus surface to the base. The base is an equilateral triangle. The lines forming its side are the plan views of the binary systems. The corners, then, represent the pure components. This method, developed by Roozeboom, divides the sides of the equilateral triangle into 100 parts. The percentage composition of each of the three components, forming the ternary system, is obtained from its position and the distance of the point P (see Fig. 9) from the three sides of the triangle in directions parallel to the sides.

The procedure of representing a three-component mixture or solution by a point within the triangle can best be shown by an illustration. Let us suppose we wish to represent a mixture composed of 50 per cent of L, 30 per cent of M, and 20 per cent of N. First 50 units are measured off on the side of the triangle LM (Fig. 9) from the corner M. Let this be point A. Next the constructional line AB is drawn parallel with the line MN, as shown. From L, 30 units are measured off on the side ML. This is point C. From C the line CD is drawn parallel to LN.

AB and *CD* intersect at point *P*. Through *P* the line *EF* is drawn parallel to the remaining side, *ML*. The point *P* represents the composition of this ternary system. By consulting the figures at the side of the triangle the method of representation is made clear.

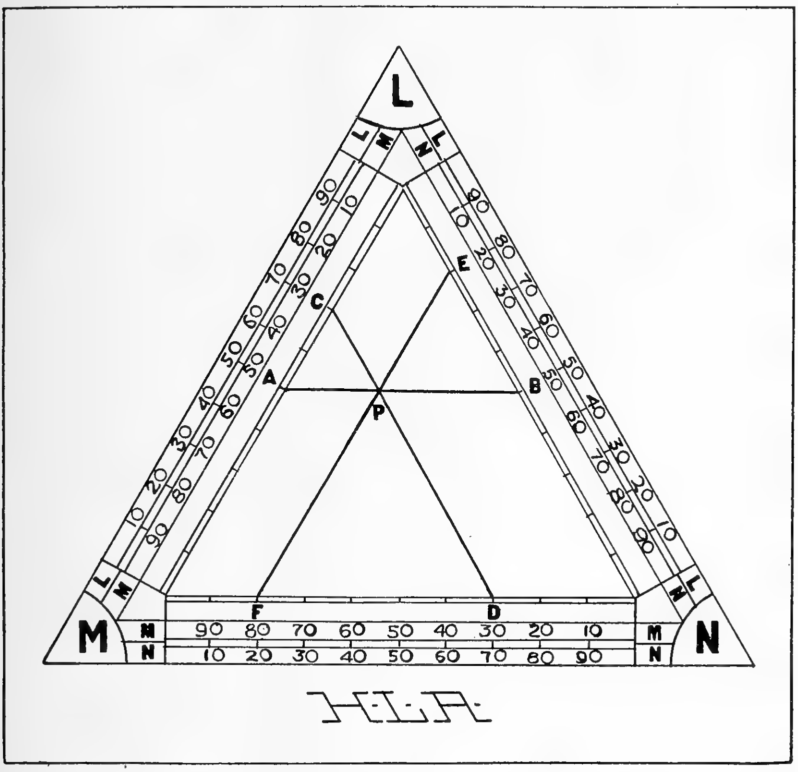


FIG. 9.—Ternary diagram, illustrating the method of indicating by a point (P) the composition of a specimen consisting of 50 per cent of *L*, 30 per cent of *M*, and 20 per cent of *N*.

By means of such triangular diagrams we can represent in a most direct manner the thermo-equilibrium diagram of a three-component system. If we employ a sufficient number of these figures we could discuss the complete feldspar system with the aid of the phase rule. The following are the ternary systems that would be necessary to that end: (1) potash-soda-lime, (2) potash-soda-barium, (3) potash-soda-carnegieite, (4) soda-lime-barium,

(5) soda-lime-carnegieite, (6) soda-barium-carnegieite, (7) lime-barium-carnegieite, (8) potash-barium-carnegieite, (9) potash-barium-lime, and (10) potash-lime-carnegieite.

As it is not possible, within the limits of the present paper, to discuss all of these ternary systems, we are compelled to confine our considerations to the most important series, the potash-soda-lime feldspars.

THE POTASH-SODA-LIME FELDSPARS

The first important paper outlining the ternary system, K-, Na-, Ca-feldspars, appeared in 1905 from the pen of J. H. L. Vogt.[1] Tracings of his original figures are here reproduced in Figure 10. The two upper diagrams illustrate the conventional method of representing the space model. The plagioclase series will be recognized as occupying the back plane. It will be seen that Vogt's conception of the *solidus* $T_{An}dT_{Ab}$ was that it assumed a straight line, which recent laboratory work has modified to a concave one as will be recalled by consulting Figure 2. The binary solubility lines extending from the eutectic temperatures to the base of the space model, which indicate low or normal temperatures, are not drawn. In the plan view the ternary solubility lines, *hg* and *ki*, at eutectic temperature, are shown projected to the base. The plan view may represent the ternary diagram at eutectic temperatures, or the diagram at normal temperatures provided the binary solubility lines are vertical. Vogt's original conception was that these lines were vertical. However, Warren has correctly suggested that these lines should be inclined, approaching the sides of the binary diagram with lowering temperature. This modification is called for in order to explain the formation of perthites ("perthoids") due to exsolution.

The Or-An binary is more or less hypothetical.

If the plan view be considered as a transverse section of the diagram cut at eutectic temperature, the areas *AnAbgh* and *kiOr* represent solid solutions of the components at this temperature. At normal temperature these areas should be more restricted than is actually shown by Vogt, occupying less space on the diagram. Positions outside of these areas represent compositions where

[1] J. H. L. Vogt, *Tschermak's Mineralog. und Petrogr. Mitt.* (1905), pp. 24 *et seq.*

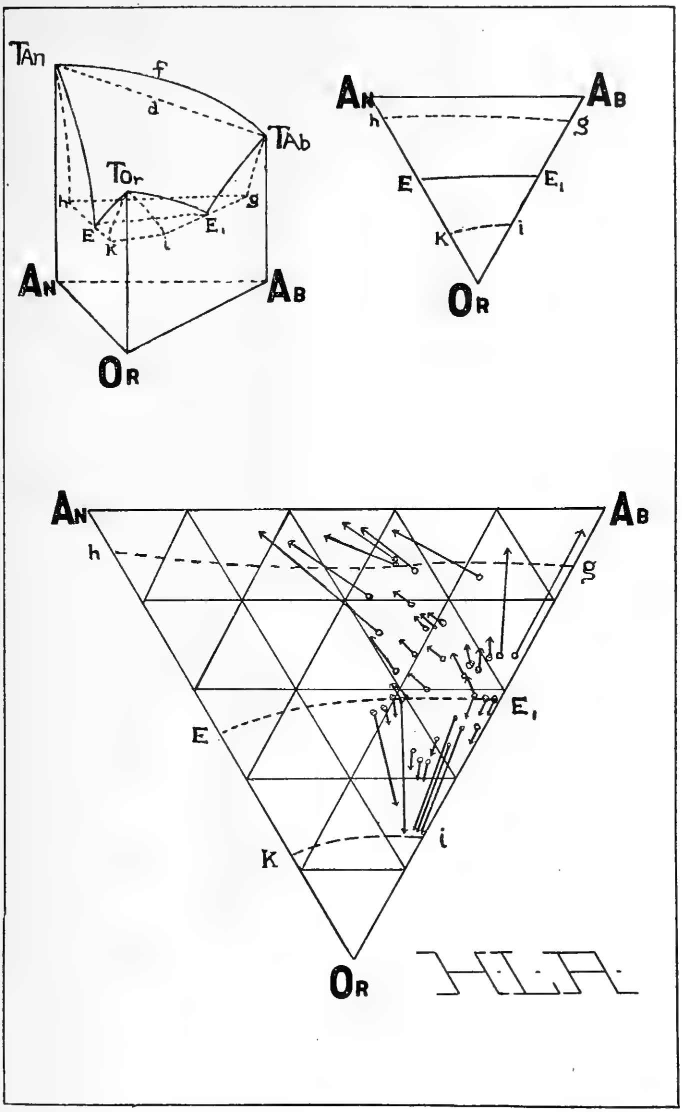


FIG. 10.—The ternary diagram of the potash-soda-lime feldspar, after Vogt. The first diagram is the perspective method of representing the solid model. The one to the right is the projection of the same to the base. The lower diagram shows the change in composition during crystallization. The circles represent the composition of the first-formed feldspars, the arrows show the direction of the change in composition taken by the subsequent forming minerals, and the arrow heads themselves indicate the composition of the last-formed minerals.

two-phase systems coexist in equilibrium. Because of the great viscosity of the high alkali feldspars we can expect that solid solutions can exist as metastable systems in the regions between the ternary solubility lines, but with increased lime content, regions near An, the ability of these feldspars to undercool and remain homogeneous at normal temperatures decreases to a point where intergrowths are rare.

The line EE_1 joins the eutectic points of the two eutectiferous binaries; and is consequently called the eutectic line. Although Vogt in the perspective drawing shows the eutectic line EE_1 horizontal, the probability is that E is situated at a higher temperature than E_1. We do not know the actual temperatures of the binary eutectics, but there is some evidence to suspect that the main portion of the solidus surface slopes down from E to E_1.

Figure 11 shows the liquidus and the solidus surfaces of the ternary system projected to the base and represented by contours of equal temperatures: isotherms. These diagrams are only approximate, but in spite of that fact it must be remembered that they constitute the rather meager basis upon which our knowledge rests.

Vogt pointed out that the feldspars which first crystallize from a magma are of different composition from those that form during the later stages of freezing. The early formed crystals are of the orthoclastic or microclinic type if the ratio of the potash component to the soda member plus anorthite was greater than the eutectic ratio. If the ratio, on the other hand, was less than that of the eutectic then the feldspars formed during the later stages of crystallization would be plagioclase. These phenomena have been indicated in Vogt's diagram in the lower part of Figure 10. The little circles represent the composition of the first-formed feldspars, and the arrows show the direction of the change in composition taken by the subsequent forming minerals, the arrow heads themselves indicating the composition of the last-formed minerals. The diagram clearly shows that during the freezing of a melt of eutectic composition the feldspars separate from that ratio toward both the potash and plagioclase areas. In other words the feldspars "split along the eutectic line." The consequences that

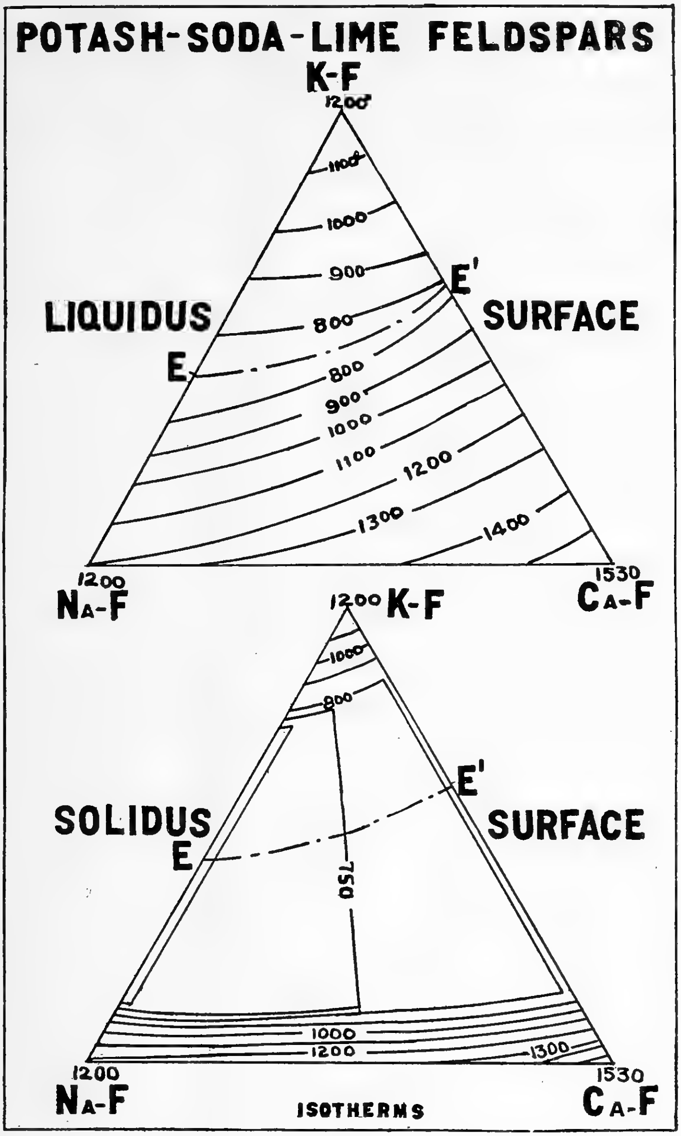


FIG. 11.—The liquidus and solidus surfaces of the ternary system, potash-soda-lime feldspars, projected to the base and represented by contours of equal temperatures or isotherms.

follow from this phenomenon are that, in the simple freezing of a magma, the maximum number of stable feldspathic phases is two, and these possess fairly constant compositions. The feldspathic content of igneous rocks, therefore, is comparatively simple in contrast with its complexity generally found in sedimentary rocks. This fact constitutes a useful criterion in distinguishing orthogneisses from paragneisses.

The complete thermo-equilibrium space model of these feldspars should show the dimorphism of the potash and the soda components. The discussion of this phase of the subject has shown that we do not possess sufficient data to supply the lines, the surfaces, and the spaces within the model to make it complete. The lack of this information seriously handicaps the interpretative petrologist. It is hoped, in spite of the obvious difficulties, that laboratory experimental work may supply the missing information.

Physical Properties of the Potash-Soda-Lime Feldspars.—In dealing with a two-component system the change in the physical properties and their relations to the change in composition is best indicated by lines upon a plane surface. In the case of three components, as here under consideration, these variables are represented by surfaces. The most convenient method of indicating to the eye the nature of these surfaces is by the use of contours. On these the *liquidus* and the *solidus* surfaces may be shown by lines of equal temperature or isotherms, the value of the extinction angles by isogonic lines, and the values of the specific gravity by lines indicating equal density, etc. All of these contoured surfaces are drawn upon triangular bases showing the interrelationships of these properties to the composition itself. To be able to show with accuracy properties of the potash-soda-lime feldspars by means of these surfaces is an ideal not yet fully realized. If the reader will bear in mind that the following diagrams are conjectural because of our total lack of definite information about the potash-lime feldspars and to be studied as the stratigrapher studies and interprets his paleogeographic maps then there will be no misunderstanding in regard to them. The writer feels that there is great value in the construction of these tentative diagrams, for

they demonstrate, as no other method can, the nature of these minerals and the constants by means of which they are identified.

Extinction Angles: Inasmuch as the extinction angles of the plagioclase and the potash-soda series constitute the most serviceable means of identification, it is important that some attempt be made to draw the isogonic lines of the orthoclase-albite-anorthite and the microcline-albite-anorthite systems for the two faces (010) and (001). This involves the use of four triangular diagrams, which are reproduced as diagrams 1 to 4 in Figure 12. When the extinction angle of natural specimen has been measured an inspection of the proper diagram shows that considerable compositional range is indicated by its curve. The extinction angle, therefore, does not appear to constitute a conclusive identification of composition. The common procedure is to ignore the least-abundant component and thus attempt to reduce the system to a simple binary. The writer is convinced, however, that this sacrifices considerable accuracy. It is much more accurate to assume the presence of a small amount of the third component rather than to ignore it altogether. As a result of examining nearly 1,300 analyses of feldspars the following empirical rule is offered: Albites contain an average of 6 per cent of the potash component; oligoclases, 8 per cent; andesines, 7 per cent; labradorites, 6 per cent; bytownites, 4 per cent; and anorthites, 3 per cent. Even though these figures are not constant (see "potash oligoclase" for example) yet the writer is convinced that if these are assumed the petrologist will be nearer to the truth than if the third component is ignored. A similar set of figures might be set up for the potash-soda side of the triangle. In soda orthoclase and soda microcline the percentage of the lime component is 2 or 3 per cent. In potash albite and anorthoclases of hyperperthitic composition the percentage of the lime component is greater, reaching in some instances as high a value as 15 or 18 per cent. This is suggested in diagram 5 in Figure 12 by the irregular line within the triangle. It represents the approximate average composition of 954 natural feldspars actually recast and plotted.

The point of intersection of this compositional line and of the measured extinction angle line represents the true composition of

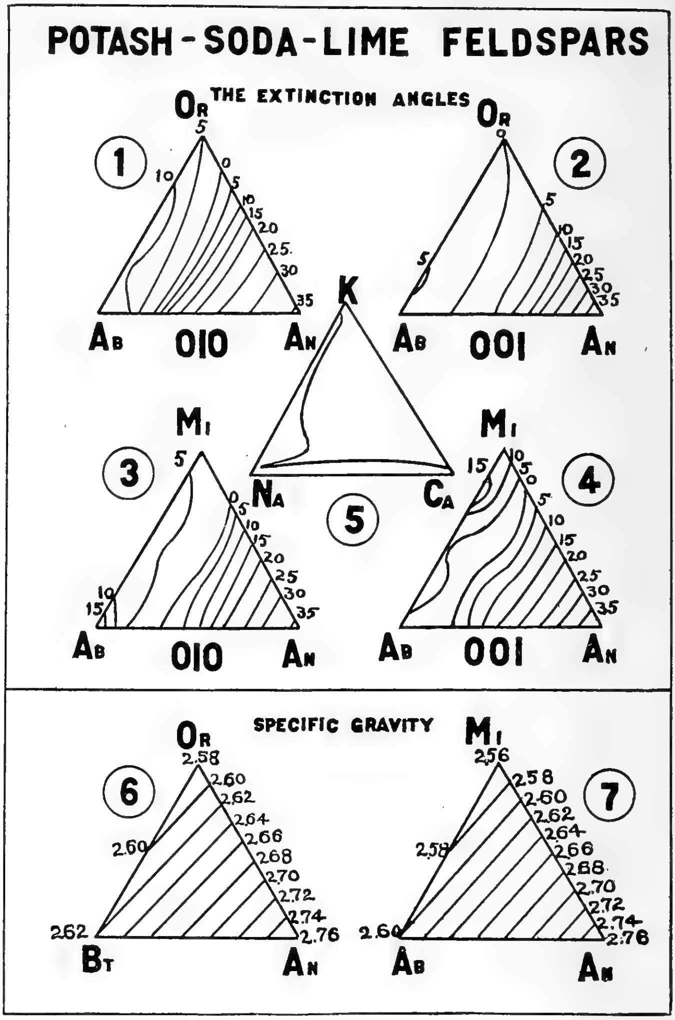


FIG. 12.—The diagrams, 1 to 4, are attempts to show the extinction angles of the potash-soda-lime feldspars by means of isogenic lines. Diagram 5 shows the approximate average composition of natural feldspars. The intersection of this compositional line and the proper isogenic lines determines the composition of the specimen under examination. Diagrams 6 and 7 show the specific gravities of the orthoclase-barbierite-anorthite and the microcline-albite-anorthite systems. All of these diagrams are tentative.

the specimen more accurately than the intersection of the isogonic line and the side of the triangle. The latter procedure is the one commonly followed although the petrologist does not express the method in these terms.

An important fact, which needs to be emphasized, is that in using the conventional extinction curves of the textbooks to determine the percentage of the soda component in a plagioclase, it is not possible to determine the percentage of the lime and potash members with anywhere near the same accuracy. The reason for this fact is that the isogonic lines are nearly parallel to the potash-soda side of the triangle. It follows then that the most satisfactory means of determining the true composition, the percentage of each of the three components, is to consider that the value of the extinction angles gives the percentage of the soda component only. Knowing this fact it can be ascertained what particular subdivision of the series is being examined. By consulting the compositional line or by reference to the average percentage of the potash component in the natural plagioclase, the approximate amount of the third component is found. The amount of the line component is the remainder. This can be illustrated: Suppose the extinction angles of a specimen were found to be 23° on the (010) face and 10° on the (001) face. The theoretical composition of this specimen would be found from the plot shown in Figure 3 to be $Ab_{40}An_{60}$, but by the method here suggested the value of the Ab alone is correct. Now it has been found that most labradorites contain an average of 6 per cent of the potash component. Thus the amount of the lime member is 60 minus 6, or 54 per cent. Therefore the composition of the specimen is K-feldspar, 6; Na-feldspar, 40; Ca-feldspar, 54.

The fact that the isogonic lines are nearly parallel to the potash-soda side of the triangle is the reason why it is easier to determine the composition of the plagioclase feldspars than that of the potash-soda series; for the isogonic lines intersect that side of the diagram more frequently. To state it in another way, there is a greater change in the value of the extinction angles per unit change in composition. As most petrographic determinations of extinction angles are only approximate it follows that an error in

the extinction angles of the plagioclase feldspars does not involve such a large error in the determination of the composition as it does in the case of the potash-soda feldspars. This is, perhaps, the reason that petrographers have thought it was impossible to determine, microscopically, the chemical composition of the latter series.

Specific Gravities: We have seen that the specific-gravity curves for the potash-soda series probably are represented by two non-parallel lines; the upper lines representing the density of the monoclinic modifications and the lower those of the triclinic forms. The specific gravities of the plagioclase feldspars, Ab-An, can properly be represented by a straight line inclined upward from albite to anorthite. Taking the curves for these binary systems it has been possible to construct the diagrams 6 and 7 in Figure 12 for the ternary system. They are instructive, even though they lack confirmation on the potash-lime side. They also illustrate the difficulty in the determination of the potash-soda series in that only a few lines intersect the potash-soda binary side, while a greater number cut the plagioclase side.

The diagrams indicate an inclined flat surface of the space model which they represent. One diagram is for the orthoclase-barbierite-anorthite system; the other, the microcline-albite-anorthite feldspars. Two others might be drawn: orthoclase-albite-anorthite and microcline-barbierite-anorthite.

CLASSIFICATION

It is important that we attempt to sum up the ternary system, potash-soda-lime feldspars, by offering a reasonable classification of the same. To accomplish this the writer has had to venture upon untrodden ground and therefore realizes his limitations. To overcome some of the objections that may be raised against it on the ground that it is too comprehensive or too complicated, the scheme is offered in two forms: one which may be called the technical classification and the other the popular one. They are shown in Figure 13.

In proposing these classifications absolutely new names have been avoided so far as possible. For the row of areas beginning with potash oligoclase and ending with potash bytownite the

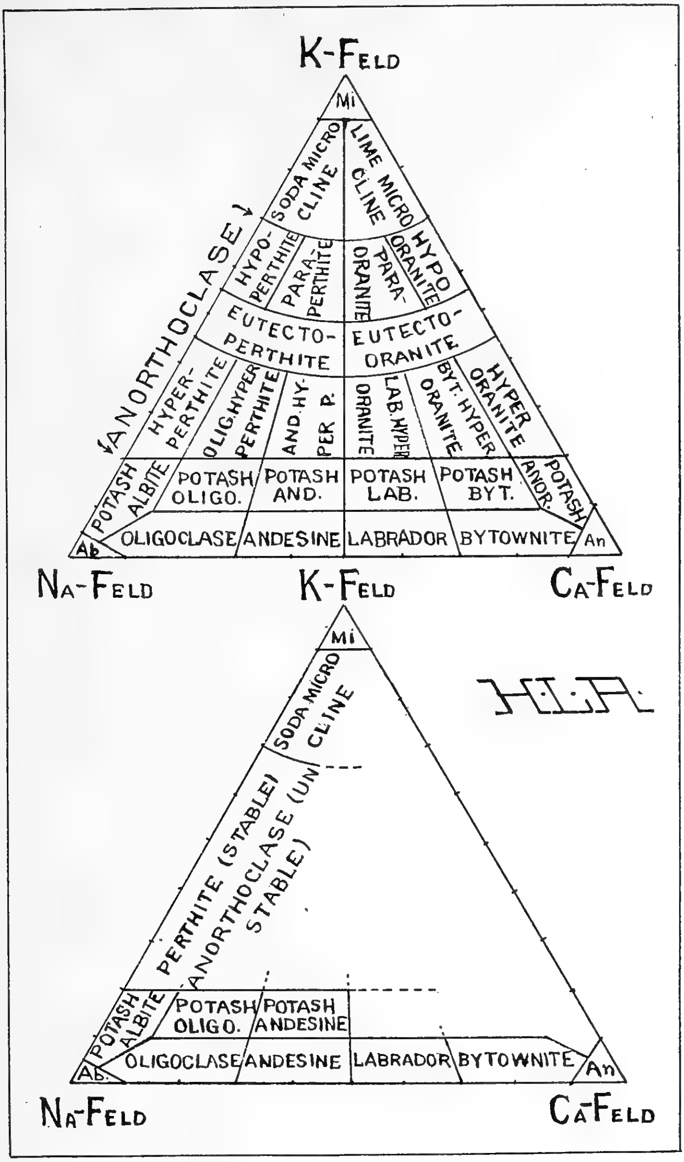


FIG. 13.—Proposed schemes for classification of the potash-soda-lime feldspars. The upper diagram is the "technical" while the lower one is the "popular" scheme. Orthoclasic feldspars are not indicated; the microclinic forms being shown instead.

writer has merely extended the term "potash" as given to the feldspar from Tyveholmen, Norway, to other plagioclasic feldspars with a potash-component content greater than 10 per cent. The prefixes hypo-, eutecto-, and hyper- have been pointed out before as adaptations of metallographic nomenclature. The term "oranite" is employed for feldspars of the potash-lime ranges analogous to perthite. The derivation of the term is indicated by the first two letters of *or*thoclase and *an*orthite with the mineralogical ending *-ite*. Thus with these prefixes and the term "oranite" many areas can be supplied with suitable names. In the triangles between hypoperthite and hypo-oranite the proposed terms "para-perthite" and "para-oranite" are not very satisfactory. The Greek *παρα-* possesses a wide range of meanings, one of which implies "a position alongside of" and "beside of," which is the meaning desired here. Criticism of the term, however, may be made on the ground that the geologist and the mineralogist use paragneiss, paraschist, paragenesis, paramorphism, paramorph, etc., without attaching this significance to the prefix. It is a term which can be temporarily used until a better one is found.

EXAMINATION OF CHEMICAL ANALYSES OF FELDSPARS

It is very desirable to know what is the actual composition of natural specimens of feldspars, and how these have been classified. Consequently many chemical analyses have been examined. About 1,300 analyses of feldspars from all parts of the world were collected from the literature. Of this number 954 were considered to be suitable for recasting and plotting. In dealing with such a large number of analyses it was obviously impossible to recast each by first obtaining the molecular ratio of each oxide and combining them in the usual manner. Therefore short cuts to approximately the same results were used. It was first assumed that each specimen was composed of only three components, the potash, soda, and lime feldspars, excepting in case where BaO was determined, indicating the presence of celsian.

The percentage of each component was determined directly from the percentage of the characteristic base. Thus all of the K_2O was assumed to be in $KAlSi_3O_8$; all the Na_2O in $NaAlSi_3O_8$, and

the CaO in $CaAl_2Si_2O_8$. The small amounts of extraneous bases and water, such as Fe_2O_3, FeO, MgO, H_2O, etc., that are very frequently present in natural specimens, have been ignored. The following factors were used in calculating the percentage of each of the three components:

Component	Formula	Oxide	Factor
Potash........	$KAlSi_3O_8$	K_2O	16.85
Sodium.......	$NaAlSi_3O_8$	Na_2O	11.83
Lime.........	$CaAl_2Si_2O_8$	CaO	20.21

The sum of these three components should be 100 per cent theoretically but as a matter of fact it rarely was. The majority of these totals was about 96 per cent, indicating if the analyses were accurately made, that the specimens were only 96 per cent pure feldspar. The inferior quality of many of the analyses or the probable presence of additional components in the system were emphasized when the sum of the feldspar components was considerably above or below 100. All those showing a total below 85 and above 110 have been rejected. The three feldspar components were then proportionately raised or lowered to 100 per cent. All the calculations were performed on a 20-inch slide rule; the errors resulting from its use being within the limits of the chemical analyses and within plotting range upon the triangular co-ordinate base used (Fig. 19).

It is known that not all of the Na_2O is necessarily in $NaAlSi_3O_8$ in every case. This base may be present in the form of nephelite or carnegieite in addition to the albite. Foot and Bradley[1] have pointed out such a possibility and say: "Albite sometimes occurs, associated with an excess of the constituents . . . . either in free condition, as corundum or silica, or in combination as nephelite." When the sum of the three feldspar components, calculated from a reliable chemical analysis, exceeds 100 per cent then the number of components is probably in excess of the three assumed. This condition is far more common than is usually supposed. Washington

[1] Foot and Bradley, "On Solid Solutions in Minerals, III," "The Constant Composition of Albite," *Amer. Jour. Sci.* (4), XXXVI, 47.

and Wright[1] have shown that the plagioclase from Limosa contains an appreciable amount of carnegieite in solid solution.

The chemical analysis is as follows:

SiO_2	52.77	CaO	10.66
Al_2O_3	29.50	Na_2O	5.40
Fe_2O_3	.65	K_2O	.74
FeO	.17	H_2O	.36
MgO	.05	Total	100.30

They recast the analyses in terms of four components:

Potash component	$KAlSi_3O_8$	4.48
Sodium component	$NaAlSi_3O_8$	36.16
Lime component	$CaAl_2Si_2O_8$	53.78
Carnegieite	$Na_2Al_2Si_2O_8$	5.58

If, however, the presence of the carnegieite in the feldspar was not suspected and the analysis recast on the basis of only the three components then the result would be as follows:

Potash component	4.48
Sodium component	54.20
Lime component	53.78

with the excessive total of 112.46 per cent. A short method of obtaining the proper composition of a carnegieite bearing feldspar can be secured by setting up the ratio, Na-component without carnegieite: 1 = Na-component with carnegieite: .6667.

Because of the probability that many of the analyses here recast and plotted, even though tested by the method above mentioned, are inferior to those now being made in many laboratories, there are limitations to the conclusions that can be safely drawn from their study and comparison. Yet it is believed that they illustrate beyond much doubt that the term "orthoclase" is used in a very loose manner, quite inconsistent with present-day standards. Fair maximum and minimum limits for the range assumed by natural "orthoclase" specimens among the analyses examined were:

	K-feldspar	Na-feldspar	Ca-feldspar
from	87.60	11.05	1.35
to	49.20	48.60	2.20

[1] H. S. Washington and F. E. Wright, "A Feldspar from Limosa and the Existence of a Soda-Anorthite (Carnegieite)," *Amer. Jour. Sci.* (4), XXIX (1910), 52–70.

Furthermore in only a very few cases where the three oxides, Na_2O, K_2O, and especially CaO, had been looked for, was any one of the three components entirely wanting. Such a revelation may not be surprising to the petrographer for he knows that it is usually impossible to determine with a microscope the composition of the alkali feldspars with anything like the accuracy obtainable in case of the plagioclase series.

The writer has secured some sixty specimens of so-called orthoclase from many world-famous localities and has satisfied himself after careful petrographic examination of them in thin sections and in crushed fragments that in the majority of cases the mineral is not orthoclase at all, but that it is a microcline relatively high in soda and more frequently a microclinic perthite. It is but a reasonable assumption, therefore, that the specimens which furnished the material from which the analyses have been made had not been examined petrographically, for if such examinations had been made, the name orthoclase would not have been applied to them in such a careless manner. The names given in Figure 19 are the original ones published in connection with the chemical analyses which, as already stated, have been secured from many sources. The works on mineralogy by Dana, Hintze, Bayley, etc., have contributed many. Various bulletins of the state and federal geological surveys have been consulted. The volume of the Asches on *The Silicates in Chemistry and Commerce* has furnished a considerable number of analyses of the plagioclase series. The analyses themselves are not here reproduced but the references to the literature are given in the bibliography.

Hintze does not distinguish the analyses of orthoclase from those of microcline, grouping them together. This necessitates the symbol for "orthoclase and microcline." The purpose of this triangle diagram did not warrant an extreme degree of accuracy and consequently the circles have been located as close to the actual recast figures as possible without any overlapping, which would cause undue confusion in recognizing the different species there represented.

In many ways the diagram (Fig. 19) speaks for itself. It clearly indicates that most if not nearly all feldspar specimens are three-component systems.

MICROSCOPIC EXAMINATION OF NATURAL FELDSPARS

The present section contains the results of microscopic and chemical analyses of typical feldspars from many parts of the world.

The optical properties of the feldspars were only determined to the extent necessary for an identification of the species. The extinction angles on (010) and (001) were determined by Michel-Levy's "statistical" method[1] on crushed fragments, which were properly sized by passing through sieves of 100 mesh, caught on screens of 120 mesh, and then placed in a suitable mounting fluid —such as clove oil. A large share of the determinations were made with monochromatic (sodium) light and the stage of the petrographic microscope rotated from each position to extinction. The recorded results are the averages of five readings. The orientation of the fragments was ascertained in the following manner. The faces were recognized by the cleavage, the interference figure, and the twinning. The fragments broken into plates parallel to the base (001) were found to be more common than those parallel to (010); the former were often identified by traces of the albite twinning. The (010) faces are more likely to have parallel edges due to the cleavage.

For this experimental work the writer naturally chose specimens of feldspars whose chemical composition was known, and whose chemical analyses were later recast into percentages of the three feldspar components. Upon this basis the interrelationship of the optical properties and the composition were ascertained which enabled the writer to draw the diagrams of the extinction angles of the system. The reader must remember that the close check between the chemical analyses and that inferred from the optical characters is the result of using the chemical analyses for the purpose of establishing the relations between chemical composition and physical properties, and so lead to methods of identification.

The percentage of the feldspar phases, when more than one was present in the specimens, was determined by a method analogous

[1] Michel-Levy, "De l'emploi du microscope polarisant a lumière parallele pour l'étude des plaques minces les roches eruptives," *Ann. des Mines* (December, 1877), pp. 392, 471.

to that developed by Rosiwal. The outlines of the different phases in a given field of the thin section were traced upon paper with the aid of a camera lucida. The areas of these grains were measured by a polar planimeter. The sum of the areas occupied by the grains of the different minerals was assumed to be proportional to their volumes. By multiplying the volumes by the specific gravities of the minerals the proportion by weight was secured and then calculated to 100 per cent. Usually four different microscopic fields in each slide were analyzed and their results averaged. Care was taken to use an optical system (objectives and oculars) which would give the largest practical field. The composition of each phase was determined by the extinction angles of crushed fragments. The first portion of each table shows the percentage of each of the phases present and their composition. The composition of each phase is given in percentages of each of the three feldspar components, totaling 100 per cent. In the second portion of the tables the percentage of each component is calculated upon a basis of 100 per cent for the entire specimen. The sum of the different components thus obtained gives the composition of the entire specimen after the manner of a recast chemical analysis. Consequently a chemical analysis, recast, would have to be rearranged by distributing the components into the various phases present in order to appreciate the true nature of the specimen. This indicates that the perthitic feldspars are much more complicated than is generally thought. The accuracy of the proper distribution of the components into the phases is directly dependent upon the accuracy of the thermo-equilibrium diagram and the degree of undercooling of the feldspar system under consideration. Until the diagrams of these minerals can be put upon a quantitative basis our examinations will be approximate only.

EXAMPLES OF PLAGIOCLASE FELDSPARS

1. "Albite," Amelia Court House, Virginia. (Specimen 961.)

Microscopic examination, thin section: Broad albite twinning.

Extinction angles, crushed fragments:

(010) 18.3°
(001) 3.0°

Inferred composition:

K-feldspar........ 1.0
Na-feldspar....... 97.0
Ca-feldspar........ 2.0

CHEMICAL AND MICROSCOPIC ANALYSES

	SiO_2	Al_2O_3	CaO	MgO	Na_2O	K_2O	H_2O	Fe_2O_3	Total
1.........	67.06	21.72	1.59	.03	10.01	.39	......	......	100.80
2.........	68.44	19.35	....	....	11.67	.43	......	......	99.89
3.........	68.22	19.06	.40	....	11.47	.20	.69	.15	100.19

	K-feldspar	Na-feldspar	Ca-feldspar
1.........	2.24	90.60	7.14
2.........	2.49	97.51	....
3.........	1.2	96.9	1.9
Microscopic	1.0	97.0	2.0

Classification: Albite.

Analyses 1 and 2. Foote and Bradley, *Amer. Jour. Sci.* (4), XXXVI (1913), 47, after Robertson and Musgrave. *Chem. News,* XLVI (1882), 204. Dana, *System of Mineralogy.* Albite 15.

Analysis 3. Allen and Day, *Carnegie Inst. Pub. 31,* p. 48, G. P. Merrill.

2. "Oligoclase," Arendal, Norway. (Specimen 960.)

Microscopic examination, thin section: Perfectly normal plagioclase with albite twinning.

Extinction angles, crushed fragments:

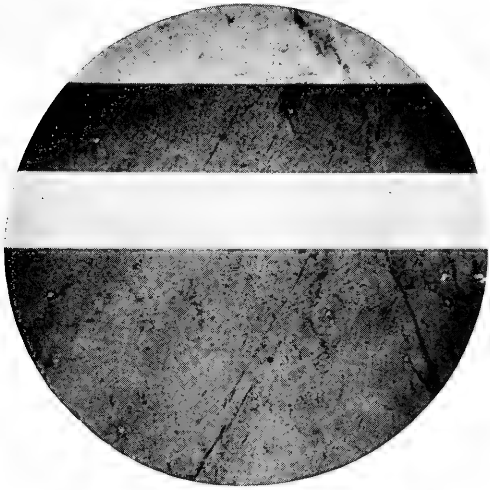
(010) 4.0°
(001) 1.0°

Index of refraction, beta: 1.547

Inferred composition, ignoring potash component $Ab_{75}An_{25}$.

Inferred composition, considering potash component:

K-feldspar........ 10.0
Na-feldspar....... 75.0
Ca-feldspar....... 15.0



A

B

A. Albite, Amelia Court House, Virginia. Polarized light. ×30. Specimen 961.

B. Labradorite, near Nain, Labrador. Polarized light. ×30. Specimen 967.

CHEMICAL AND MICROSCOPIC ANALYSES

	SiO_2	Al_2O_3	CaO	K_2O	Na_2O	MgO	H_2O	Total
1	63.53	24.05	2.60	1.86	8.02	....	.90	100.00
2	63.51	23.09	2.44	2.19	9.37	.77	....	101.37

	K-feldspar	Na-feldspar	Ca-feldspar
1	11.8	74.5	13.7
2	12.3	76.0	11.7
Microscopic	10.0	75.0	15.0

Classification: Oligoclase.

Analysis 1. Asch, *The Silicates in Chemistry and Commerce.* Analyses 9 and 34, which are duplicated. Asch 9, Des Cloizeaux, *Bull. Soc. Min.* 7 (1884), 225. Asch 34, Rosales, *Pogg. Ann.* (1842), 55, 109. Analyzed by Dirvell.

Analysis 2. Asch, *ibid.*, Analysis 71, Hagen, *Pogg. Ann.* 44 (1838), 329. Analyzed by Hagen. See Dana, *System of Mineralogy.* Extinctions on:

ARENDAL

CaO Percentage	001	010
2.50	0–2	10–12.5
2.60	0–1.5	9–12
2.81	0–2	10–12
4.20	.5–1	2–4

3. *"Oligoclase, Sunstone," Tvedestrand, Norway.* (*Specimen 970.*)

Microscopic examination, thin section: Oligoclase with inclusions of hematite, which are in all probability due to exsolution.[1]

Extinction angles, crushed fragments:

	Mallard*	Schuster†	Andersen‡	Dana§	Alling
(010)	2°–5°	3°34′	3.5°	2°–4°	2.5°
(001)	1°–1°27′	1°10′	1°	1°30′	1.0°

* E. Mallard, *Bull. Soc. Min. France*, IV (1880), 104.
† Max Schuster, *Tschernaks Min. und Petrog. Mitt.*, III (1880), 164.
‡ Olaf Andersen, *Amer. Jour. Sci.* (4), XXX (1915), 379–80.
§ James D. Dana, *System of Mineralogy*, sixth ed., p. 336.

Inferred composition from the writer's measurements, ignoring the potash component: $Ab_{70}An_{30}$.

Inferred composition, considering potash component:

K-feldspar	7.0
Na-feldspar	70.0
Ca-feldspar	23.0

[1] Olaf Andersen, *Amer. Jour. Sci.* (4), XXX (1915), 379.

CHEMICAL AND MICROSCOPIC ANALYSES

	SiO_2	Al_2O_3	Fe_2O_3	CaO	K_2O	Na_2O	Total
Analysis	61.30	23.77	0.36	4.78	1.29	8.50	100.00

	K-feldspar	Na-feldspar	Ca-feldspar
Chemical	7.45	69.62	22.93
Microscopic	7.0	70.0	23.0

Classification: Oligoclase.

Analysis. James D. Dana, *System of Mineralogy.* Oligoclase and oligoclase-albite No. 11. Asch, *The Silicates in Chemistry and Commerce.* No. 40. Scheerer, *Poggendorff's Annalen,* 64, 1845, 153.

4. *"Labradorite," near Nain, Labrador.* (*Specimen 967.*)

Microscopic examination, thin section: Normal plagioclase with albite twinning. Small inclusions of rutile, ilmenite (?), diopside, and hematite. Extinction angles, crushed fragments:

	(010)	20°
	(001)	12°
Maximum, Zone ⊥	(010)	33°

Inferred composition, ignoring potash component: $Ab_{44}An_{56}$.
Inferred composition, considering potash component:

K-feldspar	2.0
Na-feldspar	44.0
Ca-feldspar	54.0

CHEMICAL AND MICROSCOPIC ANALYSES

	SiO_2	Al_2O_3	CaO	Na_2O	K_2O	Ign	Fe_2O_3	MgO	Total
1	56.00	27.50	10.10	5.00	.40	......	.70	.10	99.80
2	56.18	27.33	10.33	5.17	.36	......	1.38	......	100.75
3	54.75	27.76	10.60	5.13	.53	.56	.69	......	100.02

	K-feldspar	Na-feldspar	Ca-feldspar
1	2.50	44.70	52.41
2	2.21	45.10	52.61
3	3.18	43.80	53.01
Microscopic	2.0	44.0	54.0

Classification: Labradorite.

Analyses. James D. Dana, *System of Mineralogy, Labradorite Nos. 23, 24, 25.*

EXAMPLES OF POTASH-SODA FELDSPARS

1. "Orthoclase," "Adularia coated with chlorite," Scopi, Switzerland. (Specimen 988.)

Microscopic examination, thin section: Clear, untwinned potassic feldspar. Extinction angles, crushed fragments:

(010)	5.0°
(001)	0.4°

Inferred composition, considering potash component:

K-feldspar..........	92.0
Na-feldspar.........	6.0
Ca-feldspar..........	2.0

This is one of the very few specimens examined to which the term "orthoclase" can be assigned in accordance with the nomenclature here adopted. Classification: True orthoclase.

2. "Chesterlite," Poor House Quarry, West Chester, Pa. (Specimen 997.)

Microscopic examination, thin section: Broad phantom twinning, suggesting microcline.
Extinction angles, crushed fragments:

(010)	6.0°
(001)	16.5°

Inferred composition, considering the potash component:

K-feldspar..........	83.0
Na-feldspar.........	14.0
Ca-feldspar..........	3.0

CHEMICAL AND MICROSCOPIC ANALYSES

	SiO_2	Al_2O_3	K_2O	Na_2O	Ign	Fe_2O	CaO	MgO	Total
1...........	64.97	17.65	14.02	1.69	.65	.50	.50	.27	100.36

	K-feldspar	Na-feldspar	Ca-feldspar
Chemical....	82.8	14.2	3.0
Microscopic..	83.0	14.0	3.0

Classification: Popular and technical: Soda microcline.

Analysis. James D. Dana, *System of Mineralogy.* Microcline 5. Hintze, *Handbuch der Mineralogie.* "Orthoclase and Microcline Kalifeldspath," CCXCII.

3. *"Orthoclase, var. Adularia," Eggerhorn, Switzerland. (Specimen 989.)*

Microscopic examination, thin section: Very coarse phantom twinning with wavy extinction. Some areas comparatively free from microclinic texture, and appear like perfect development of orthoclasic feldspar.

Extinction angles, crushed fragments:

Untwinned	(010)	7.2
Untwinned	(001)	1.4
Twinned	(010)	6.5
Twinned	(001)	17.0

Inferred composition:

K-feldspar	82.0
Na-feldspar	16.0
Ca-feldspar	2.0

The probable character of the feldspar is that it is in the process of inverting from soda orthoclase to soda microcline.

Classification: Soda orthoclase—soda microcline.

4. *"Microcline," Georgetown, Maine. (Specimen 995.)*

Microscopic examination, thin section: It is readily seen that the specimen is a microclinic microperthite.

Extinction angles, crushed fragments:

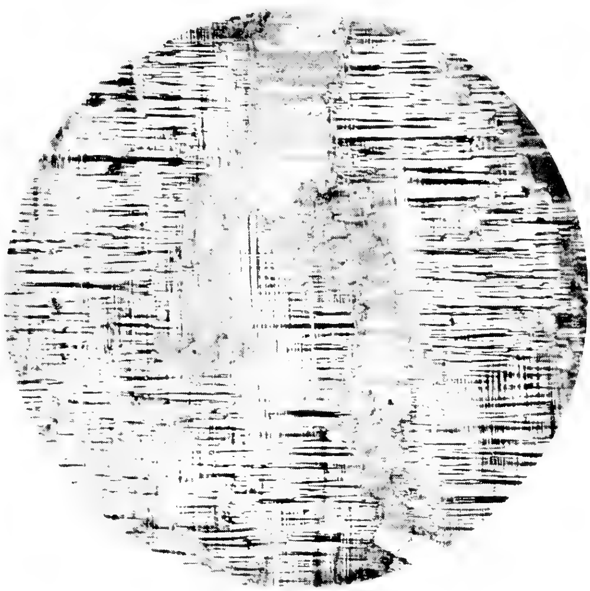
Potash phase	(010)	6.0°
	(001)	15.5°
Soda phase	(010)	4.0°
	(001)	1.0°

Inferred Composition	Potash Phase	Soda Phase
K-feldspar	88.0	15.0
Na-feldspar	5.0	75.0
Ca-feldspar	7.0	10.0

QUANTITATIVE MICROSCOPIC ANALYSIS

Phase	Percentage	K-feldspar	Na-feldspar	Ca-feldspar
Soda microcline	83.5	88.0	5.0	7.0
Oligoclase	16.5	15.0	75.0	10.0
	K-feldspar	Na-feldspar	Ca-feldspar	
Soda microcline	73.5	4.2	5.8	....
Oligoclase	2.5	12.3	1.6	....
Total	76.0	16.5	7.4	

A



B

A. Soda orthoclase inverting to soda microcline, called "Orthoclase, var. Adularia," Eggerhorn, Switzerland. Polarized light. ×30. Specimen 989.

B. Microcline microperthite (hypoperthite), called "Amazonstone," Amelia Court House, Virginia. Polarized light. ×30. Specimen 992.

CHEMICAL AND MICROSCOPIC ANALYSES

	SiO_2	Al_2O_3	Fe_2O_3	CaO	Na_2O	K_2O	Ign	Total
1	65.50	19.56	......	1.58	.55	12.67	......	99.84
2	65.23	20.09	.71	......	2.00	11.60	.36	99.99

	K-feldspar	Na-feldspar	Ca-feldspar
1	85.8	5.3	8.9
2	80.2	19.8	....
Microscopic	76.0	16.5	7.5

Classification:

Popular: Microcline microperthite

Technical: Microcline hypoperthite.

Analysis 1. Collection of A. A. Robbins, now on exhibition in the New York State Museum, Albany, New York.

Analysis 2. E. S. Bastin, *U.S. Geol. Surv. Bull. 420*, p. 24.

5. *"Microcline Amazonstone," Amelia Court House, Amelia County, Virginia. (Specimen 992.)*

Microscopic examination, thin section: Microclinic microperthite.

Some of the intergrowths of the potash and soda phases have an appearance as though they were primary, due to the freezing of the eutectic mixture.

The albite-oligoclase occurs in stringers which vary in thickness and are not continuous. The "linkage" areas between these blebs are characterized by a much finer Scotch-plaid type of twinning in the soda microcline which appears to be akin to anorthoclase. These areas then may be regarded as a supersaturated crystalline solid solution changing to microperthite by exsolution.

Extinction angles, crushed fragments:

Potash phase	(010)	6.4°
	(001)	17.1°
Soda phase	(010)	12.0°
	(001)	3.0

Inferred Composition	Potash Phase	Soda Phase
K-feldspar	80.0	2.0
Na-feldspar	18.0	88.0
Ca-feldspar	2.0	10.0

QUANTITATIVE MICROSCOPIC ANALYSIS

Phase	Percentage	K-feldspar	Na-feldspar	Ca-feldspar
Soda microcline...........	91	80	18	2
Albite-oligoclase...........	9	.2	88	10
Soda microcline...........	....	73.3	16.2	2.0
Albite-oligoclase...........	....	.1	7.7	.7
Total................	....	73.4	23.9	2.7

Classification:

Popular: Microcline microperthite

Technical: Microcline hypoperthite.

6. *"Microcline," Etta Mine, one mile south of Keystone, South Dakota. (Specimen 965.)*

Microscopic examination, thin section: Microcline microperthite with quartz, diopside, sericite, and carbonates. The sodic phase of the intergrowths is oligoclase.

Extinction angles, crushed fragments:

Potash phase........	(010)...........	6.0°
	(001)...........	17.0°
Soda phase.........	(010)...........	7.0°
	(001)...........	3.0°

Inferred Composition	Potash Phase	Soda Phase
K-feldspar...........	80.	3.
Na-feldspar..........	17.	80.
Ca-feldspar..........	3.	17.

QUANTITATIVE MICROSCOPIC ANALYSIS

Phase	Percentage	K-feldspar	Na-feldspar	Ca-feldspar
Soda microcline...........	87.5	80	17	3
Oligoclase................	12.5	3	80	17
Soda microcline...........	....	70.	14.9	2.6
Oligoclase................	....	.3	10.0	2.2
Total................	....	70.3	24.9	4.8

Classification:

Popular: Microcline microperthite.

Technical: Microcline hypoperthite.

7. *"Microcline," San Diego County, California. (Specimen 958.)*

Microscopic examination, thin section: Soda microcline, showing but faint microclinic twinning, and in a few areas, entirely clear. Slivers removed from the specimen with the gentle application of the knife blade show no twinning. The suggestion is strong that in grinding pieces for thin sections sufficient pressure was present to hasten the inversion of the soda orthoclase to soda microcline. To test this theory crushed fragments were heated in a quartz crucible over a Scimatco burner for one hour, three hours, and five hours. The percentage of the twinned specimens was measured in each case after the fragments had cooled. The results are tabulated below:

No.	Time of Heating in Hours	Number of Twinned Fragments Seen	Number of Untwinned Fragments Seen	Percentage of Twinned Fragments
1	0	28	64	19.6
2	1	53	90	36.8
3	3	56	48	54.0
4	5	81	34	65.0

Extinction angles, crushed fragments:

Twinned	(010)	6.5°
	(001)	17.0°
Untwinned	(010)	7.0°
	(001)	1.5°

These observations point to the fact that the specimen is in the act of inverting from the one modification to the other.

QUANTITATIVE MICROSCOPIC ANALYSIS

Phase	Percentage	K-feldspar	Na-feldspar	Ca-feldspar	Silica
Soda microcline	80	80	18	2	
Oligoclase	17	2	85	13	....
Quartz	3	....	....	....	100
Soda microcline	....	64.0	14.4	1.5	....
Oligoclase	....	0.3	14.5	2.2	....
Quartz	....	....	....	....	3
Total	....	64.3	28.9	3.8	3.0

Classification:

Popular: Microcline microperthite

Technical: Microcline hypoperthite.

8. *"Orthoclase," Sanidine Porphyry, Drachenfels, Siebenebirge, Rhenish Prussia. (Specimen 986.)*

Microscopic examination, thin section: The slide slows phenocrysts of sanidine, usually zonally grown, the central portion of which is slightly more potassic than the margins.

Extinction angles, crushed fragments, the average of a large number.

(010)	9.3°
(001)	6.0°

Inferred composition:

K-feldspar	62
Na-feldspar	34
Ca-feldspar	4

CHEMICAL AND MICROSCOPIC ANALYSES

	SiO_2	Al_2O_3	CaO	K_2O	Na_2O	MgO	Ign	Total
	65.87	18.53	.95	10.32	3.92	.39	.44	99.92
	K-feldspar		Na-feldspar			Ca-feldspar		
Chemical	61.6		33.7			4.7		
Microscopic	62.0		34.0			4.0		

Classification: Anorthoclase, potentially hypoperthite.

Analysis. James D. Dana, *System of Mineralogy.* Orthoclase No. 6. (Rg. Min. Ch. 1003, 1860.)

9. *"Sunstone," Delaware County, Pennsylvania. (Specimen F3-974.)*

Microscopic examination, thin section: The slide reveals that the specimen is microclinic perthite of two periods of development. The microcline intergrown with the albitic feldspar of the first generation is holding blebs of soda-rich feldspar that are clearly the result of a later development. It is believed that the albite phase of the second generation and the small flakes of hematite are due to the decrease in solubility of these constituents of the solid phase—they are due to exsolution.

Extinction angles, crushed fragments:

Potash phase	(010)	6.5°
	(001)	17.0°
Soda phase	(010)	21.2°
	(001)	4.0°

Inferred Composition	Potash Phase	Soda Phase
K-feldspar	80	2
Na-feldspar	17	97
Ca-feldspar	3	1

QUANTITATIVE MICROSCOPIC ANALYSIS

Phase	Phase	Percentage	Percentage	Phase	K-feld.	Na-feld.	Ca-feld.	Fe_2O_3
Perthite.....	K-feld.	80	93.5	Soda micro.	80	17	3	......
			6.5	Albite 2	2	97	1	......
	Na-feld.	19.5	19.5	Albite 1	2	97	1	......
Hematite....	....	.5	......	....	......	......	......	100
Perthite.....	Ka-feld.	80		Soda micro.	59.8	12.7	2.2	......
				Albite 2	.1	5.1	.1	......
	Na-feld.	19.5		Albite 1	.4	18.9	.2	......
Hematite....	....	.5		....	......	......	......	.5
Totals.....	....	....		....	60.3	36.7	2.5	.5

Classification:

Popular: Microcline microperthite

Technical: Microcline hypoperthite.

10. *"Microcline," Verona, Ontario, Canada.* (*Specimen 959.*)

Microscopic examination. The microscope shows that the specimen is really a microcline microperthite with some accessory quartz.

Extinction angles, crushed fragments:

Potash phase........	(010)...........	6.0°
	(001)...........	17.9°
Soda phase.........	(010)...........	5.5°
	(001)...........	1.5°

Inferred Composition	Potash Phase	Soda Phase
K-feldspar...........	73	2
Na-feldspar..........	25	80
Ca-feldspar..........	2	18

QUANTITATIVE MICROSCOPIC ANALYSIS

Phase	Percentage	K-feldspar	Na-feldspar	Ca-feldspar
Soda microcline...	81	73	25	2
Oligoclase........	19	2	80	18
Soda microcline...	....	59.4	20.1	1.6
Oligoclase........	....	.4	15.2	3.4
Total........	.....100.0....	59.4	35.3	5.0

Classification:

Popular: Microcline microperthite.

Technical: Microcline hypoperthite.

11. "Microcline-Amazonstone," Mineral Hill, Pennsylvania. (*Specimen 964.*)

Microscopic examination, thin section: Under the microscope it is readily seen that there are two generations of perthitic intergrowths. The potash phase, which is intergrown with a soda-rich feldspar, is in turn full of secondary blebs of a soda phase. This may be explained as due to the fact that complete separation of the two phases was accomplished at relatively high temperatures, while the subsequent separation has taken place at more moderate temperatures. The latter phenomenon is probably due to decreased solubility as determined by the slope of the solubility-saturation curve, line *AL* in Figure 4. The fact that the "secondary" blebs of potash feldspar in the original member, rich in soda, are not seen, even under high magnifications (950 diameters), testifies to the nearly vertical character of the corresponding curve, *BM* in Figure 4.

Extinction angles, crushed fragments:

Potash phase........	(010)...........	7.2°
	(001)...........	16.9°
Soda phase.........	(010)...........	19.0°
	(001)...........	4.8°

QUANTITATIVE MICROSCOPIC ANALYSIS

Phase	Phase	Percentage	Percentage	Phase	K-feldspar	Na-feldspar	Ca-feldspar	SiO_2
Perthite.....	K-feld.	80	90	Soda micro.	80.0	17.0	3.0	......
			10	Albite 2	3.0	95.0	2.0	
	Na-feld	19.6	19.6	Albite 1	3.0	95.0	2.0	
Quartz......	Quartz	.4	......	....	......	......	......	100
Perthite.....	K-feld.	80		Soda micro.	57.6	12.2	2.2	......
				Albite 2	.2	7.6	.2	
	Na-feld.	19.6		Albite 1	.6	18.6	.4	
Quartz......	Quartz...	.4		....	......	......	......	.4
Total...	....	....		....	58.4	38.4	2.8	.4

Classification:

Popular: Microcline microperthite

Technical: Microcline hypoperthite.

12. "Orthoclase," near Unionville, Chester County, Pennsylvania. (*Specimen 996.*)

Microscopic examination, thin section: Typical microcline microperthite with inclusions of diopside, quartz, and untwinned feldspar thought to be anorthoclase.

A

B

A. Microcline microperthite (hypoperthite), called "Amazonstone," Mineral Hill, Pennsylvania. Polarized light. ×30. Specimen 964.

B. Microcline microperthite (hypoperthite), called "Amazonstone," near Florissant, California. Polarized light. ×30. Specimen 971.

Extinction angles, crushed fragments:

Potash phase........	(010)...........	7.0°
	(001)...........	17.0°
Soda phase..........	(010)...........	13.0°
	(001)...........	1.0°

Inferred Composition	Potash Phase	Soda Phase
K-feldspar...........	79.0	2.0
Na-feldspar..........	19.0	89.0
Ca-feldspar..........	2.0	9.0

QUANTITATIVE MICROSCOPIC ANALYSIS

Phase	Percentage	K-feldspar	Na-feldspar	Ca-feldspar	Silica
Soda microcline....	71.5	79.0	19.0	2.0	0.
Oligoclase.........	24.1	2.0	89.0	9.0	0.
Quartz............	....	....	....	....	100
Soda microcline....	....	56.5	13.6	1.4	....
Oligoclase.........	....	.5	21.4	2.2	....
Quartz............	....	....	....	....	4.4
Total.........	....	57.0	35.0	3.6	4.4

Classification:

Popular: Microcline microperthite

Technical: Microcline hypoperthite

13. "Microcline Amazonstone," near Florissant, California. (*Specimen 971.*)

Microscopic examination, thin section: Intergrowths of potash and soda rich feldspars. Microclinic twinning beautifully shown.

Extinction angles, crushed fragments:

Potash phase........	(010)...........	5.5°
	(001)...........	17.0°
Soda phase..........	(010)...........	12.0°
	(001)...........	4.0°

Inferred Composition	Potash Phase	Soda Phase
K-feldspar...........	80	3
Na-feldspar..........	18	85
Ca-feldspar..........	2	12

QUANTITATIVE MICROSCOPIC ANALYSIS

Phase	Percentage	K-feldspar	Na-feldspar	Ca-feldspar	
Soda microcline....	69.6	80	18	2	....
Oligoclase.........	30.4	3	85	12	....
Soda microcline....	....	55.6	12.6	1.4	....
Oligoclase.........	....	.9	25.9	3.6	....
Total.........	....	56.5	38.5	5.0	100

Classification:

Popular: Microcline microperthite.

Technical: Microcline hypoperthite.

14. "Oligoclase," Eganville, Ontario, Canada. Specimen 962.)

Microscopic examination, thin section: The specimen has been incorrectly identified; it is a microcline microperthite—not oligoclase. There is some oligoclase in the specimen but it is intergrown with soda microcline. Besides the feldspar phases, biotite and quartz are present.

Extinction angles, crushed fragments:

Potash phase........	(010)...........	6.0°
	(001)...........	18.0°
Soda phase.........	(010)...........	5.0°
	(001)...........	1.0°
Zone ⊥............	(010)...........	5.0°

Index of refraction, soda phase, beta, 1.546

Inferred Composition	Potash Phase	Soda Phase
K-feldspar...........	70	8
Na-feldspar..........	28	75
Ca-feldspar..........	2	17

QUANTITATIVE MICROSCOPIC ANALYSIS

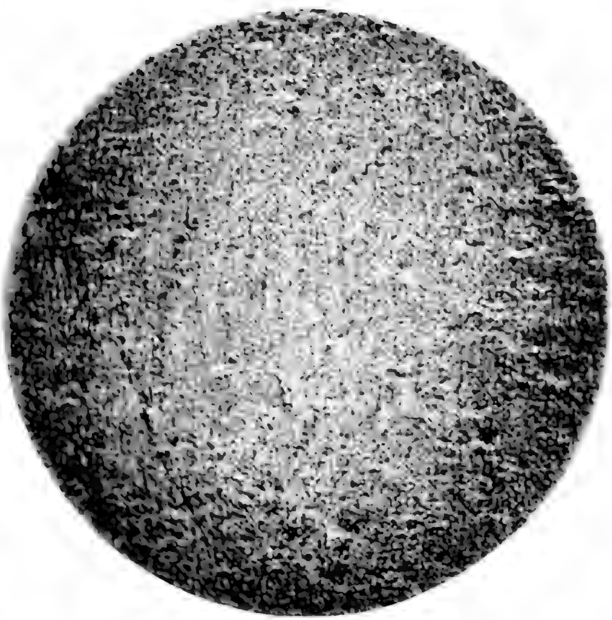
Phase	Percentage	Na-feldspar	K-feldspar	Ca-feldspar	
Soda microcline....	62	70	28	2	....
Oligoclase.........	38	8	75	17	....
Soda microcline....	....	44.4	17.3	1.2	....
Oligoclase.........	....	3.2	27.4	6.5	....
Total.........	....	47.6	44.7	7.7	100.0

Classification:

Popular: Microcline microperthite.

Technical: Microcline hypoperthite.

A



B

A. Microcline microperthite (hypoperthite), called "Oligoclase," Eganville, Ontario, Canada. Polarized light. ×30. Specimen 962.

B. Anorthoclase, Frederiksvärn, Norway. Polarized light. ×30. Specimen 981.

15. *"Anorthoclase," Frederiksvärn, Norway.* (*Specimen 981.*)

Microscopic examination, thin section: There is a faint suggestion of microclinic twinning on a very fine scale. Some areas appear to be entirely free and have slightly different optical characters. There are veinlets of oligoclase and stringers of soda microcline but they are too small for quantitative measurements.

Extinction angles, crushed fragments:

Untwinned	(010)	11.0°
	(001)	2.9°
Twinned	(010)	3.5°
	(001)	9.5°

It follows from examination of the chart, Figure 8, that the untwinned fragments are orthoclasic, that is monoclinic (?), while the twinned pieces are microclinic, triclinic (?).

Inferred composition:

K-feldspar	45
Na-feldspar	53
Ca-feldspar	2

CHEMICAL AND MICROSCOPIC ANALYSES

	SiO_2	Al_2O_3	Fe_2O_3	CaO	K_2O	Na_2O	H_2O	Total
1	65.19	19.99	.63	.48	7.03	7.08	.38	100.78
2	66.08	18.77	......	.37	7.68	6.54	......	99.44

	K-feldspar	Na-feldspar	Ca-feldspar
1	44.09	57.65	2.26
2	44.41	53.82	1.77
Microscopic	45.0	53.0	2.0

Classification: Anorthoclase, potentially eutectoperthite.[1]

16. *"Orthoclase," East DeKalb, St. Lawrence County, New York.* (*Specimen 995.*)

Microscopic examination, thin section: A typical microcline microperthite, not orthoclase. A few inclusions of apatite, calcite, phogopite, and hematite. The latter due to exsolution.

Extinction angles, crushed fragments:

Potash phase	(010)	8.0°
	(001)	11.0°
Soda phase	(010)	10.0°
	(001)	4.5°

[1] *Analyses 1 and 2.* Carl Hintze, *Handbuch der Mineralogie.* Kalifeldspath, CCXXIX and CCXXX, 1414, 1892.

Inferred Composition	Potash Phase	Soda Phase
K-feldspar...........	72	10
Na-feldspar..........	25	80
Ca-feldspar..........	3	10

QUANTITATIVE MICROSCOPIC ANALYSIS

Phase	Percentage	K-feldspar	Na-feldspar	Ca-feldspar
Soda microcline...	51.5	72.0	25.0	3.0
Oligoclase........	48.5	10.0	80.0	10.0
Soda microcline...	....	36.0	13.0	2.0
Oligoclase........	....	5.0	39.0	5.0
Total........	....	41.0	52.0	7.0

CHEMICAL AND MICROSCOPIC ANALYSES

Analysis. A. A. Robbins, Collection on exhibition in the New York State Museum, Albany, N.Y.

SiO_2	Al_2O_3	MgO	CaO	Na_2O	K_2O	H_2O	SO_3	Total
68.60	19.82	.14	.96	4.57	5.25	.30	tr	99.64

	K-feldspar	Na-feldspar	Ca-feldspar
Chemical.....	41.5	51.8	6.7
Microscopic...	41.0	52.0	7.0

Classification:

Popular: Microcline microperthite.

Technical: Microcline eutectoperthite.

17. *"Orthoclase, var. Delawarite," Lenni, Delaware County, Pennsylvania. (Specimen 991.)*

Microscopic examination, thin sections: It is evident that this specimen is what a few mineralogists call antiperthite.[1] That is, the host is a soda-rich feldspar, while the blebs or "inclusions" are high in potash.

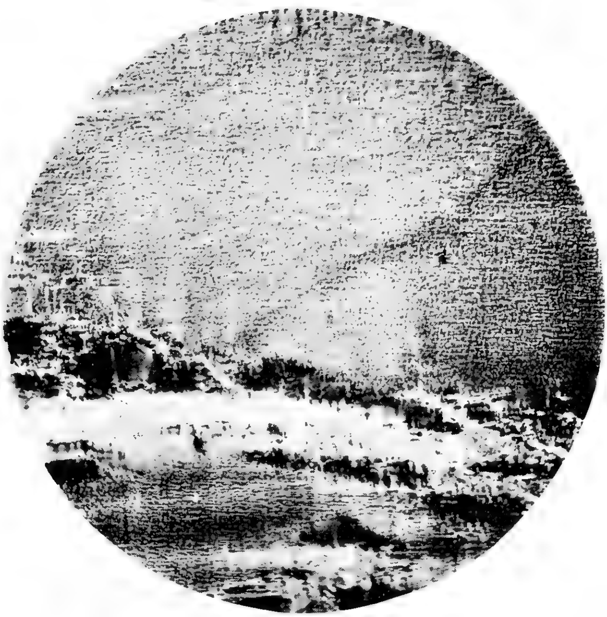
The nomenclature here adopted calls for hyperperthite. This type is rare in nature.

Extinction angles, crushed fragments:

Soda phase....	....(010)...........	14.0°
	(001)...........	4.7°
	(010)...........	5.5°
	(001)...........	16.5°

[1] Ernst Weinschenk and R. W. Clark, *Petrographic Methods* (1912), p. 326.

A



B

A. Microcline microperthite (eutectoperthite), called "Orthoclase," East DeKalb, St. Lawrence County, New York. Polarized light. ×30. Specimen 995*x*.

B. Antiperthite (hyperperthite), called "Orthoclase," var. Delawarite," Delaware County, Pennsylvania. Polarized light. ×30. Specimen 991.

QUANTITATIVE MICROSCOPIC ANALYSIS

Phase	Percentage	K-feldspar	Na-feldspar	Ca-feldspar	Silica
Potash albite	80	8	90	2	0
Primary soda microcline	10	87	10	3	0
Primary oligoclase	2	17	80	3	0
Myrmekite	2	17	75	3	5
Quartz	3	0	0	0	100
Secondary soda microcline	2	89	8	3	0
Potash albite	....	6.4	71.9	1.6	0
Primary soda microcline	....	8.7	1.0	.3	0
Primary oligoclase	....	.34	1.6	.06	0
Myrmekite	....	.34	1.5	.06	.1
Quartz	....	....	....	....	3.0
Secondary soda microcline	....	2.7	.20	.09	0
Total	....	2.7	76.30	2.11	3.1

Classification:

Popular: Antiperthite.

Technical: Hyperperthite.

APPLICATIONS OF THE MINERALOGRAPHY OF THE FELDSPARS TO GEOLOGICAL PROBLEMS

CASE ONE—LOCATION OF A FAULT

The Problem.—One of the graphite properties in the Adirondack Mountains visited by the writer in 1917[1] had been abandoned because the ore was cut off by a fault. The writer had this information when he entered the field, but to his unpleasant surprise he was unable to locate with any satisfaction the faults even though slickensided surfaces were found on the walls of the old workings.

As it was deemed very desirable to locate the faults with some degree of accuracy, the writer carried the problem into the laboratory for solution.

Method of Attack.—Preliminary examination of the graphite-bearing schist showed that it was chiefly composed of potassic feldspar. The suggestion of Rosenbusch[2] that the development of microcline structure in orthoclase is due to pressure was recalled.

[1] H. L. Alling, *New York State Museum Bull. 199*, pp. 61, 68–70, 1918.

[2] Rosenbusch-Iddings, *Microscopic Physiography of the Rock Making Minerals*, p. 320.

The writer has already discussed this phenomenon and has reached the conclusion that pressure does not *produce* microcline from orthoclase; it only starts and accelerates the change. This suggestion seemed a promising method of attack. Unless the graphite-bearing schist had suffered very severe regional metamorphism the potassic feldspar would still exist in the metastable condition which we know as orthoclase. But under the stress and jar of faulting the feldspar would take on the microclinic type of twinning as a consequence of the inversion of orthoclase to microcline. Microscopic examination would locate the fault.

Another trip into the field resulted in a collection of a suite of specimens from numerous localities in and about the old workings.

Petrographic Study.—Examination of the slides from these specimens showed that the writer's supposition was entirely correct. Some were composed of orthoclasic feldspar while others showed microclinic types.

Interpretation.—It was concluded from quantitative microscopic analyses that specimens which showed a high orthoclasic content came from areas that were free from faulting, and that specimens showing soda microcline were situated in zones affected by faulting.

Results.—In 1918, the following year, the writer took his map of the Rock Pond workings, giving the results to the petrographic study, back into the field and erected piles of stones where faulting had been deduced from the slides. From the position of these cairns it was possible to trace a group of faults that cut off the ore on three slides. Careful examination of the walls of the pits revealed conclusive evidence of the correctness of the interpretation. Slickensides and breccias were where the microscope had indicated that they should be.

CASE TWO—ORTHO-AMPHIBOLITES VERSUS PARAMPHIBOLITES

The Problem.—In many pre-Cambrian areas where ancient sediments have been invaded by igneous rocks, and subjected to contact and regional metamorphism, the character of the original rocks becomes profoundly altered, both in regard to mineralogical and structural relationships, under these forces. Both limestones and calcareous shales become metamorphosed into paraschists,

which are composed largely of hornblende. These are hornblende schists, or paramphibolites. In a similar manner old basic eruptives, such as diabases, diorites, and gabbros, are metamorphosed to orthoschists and gneisses. They may be called ortho-amphibolites. "The origin of the amphibolites is a question of the highest importance in the elucidation of the geology of the [Haliburton and Bancroft] area, as well as one of great interest from a petrographical standpoint . . . . [examination has] shown, beyond a doubt, that amphibolites, which, in many cases cannot be distinguished apart, have been produced by the action of granitic instrusions on limestone. There is also reason to believe that other amphibolites have been produced in still other ways [for there is evidence that some amphibolites are] of undoubted igneous origin."[1]

In the Adirondack Mountains ortho- and paramphibolites present a difficult problem. In areal mapping this problem is of scientific interest only, but when these are encountered upon mining properties the distinction between the two types becomes essential. The writer has encountered amphibolites[2] where it was impossible to classify the rock. In some doubtful cases the rocks were studied petrographically.

Petrographic Study and Interpretation.—Specimens were collected from rock masses where field relations pointed to a definite origin. Microscopic examinations revealed striking similarities and a few differences. The similarities need not be touched upon; it is the latter that are important. If the rock is sedimentary in origin and derived from calcareous shales as Cushing suggests,[3] quartz would be expected to occur, as unmetamorphosed shales almost universally carry some quartz. Thus if any original quartz is present in an amphibolite, it gives it a sedimentary look, for basic (subsalic, femic) rocks are usally lacking in this mineral. On the other hand the absence of quartz suggests an igneous origin,

[1] F. D. Adams and A. E. Barlow, Canada, Dept. Mines, *Geol. Surv. Mem. 6*, pp. 158–59, 1910.

[2] See J. F. Kemp and H. L. Alling, "The Geology of the Ausable Quadrangle," *New York State Mus. Bull.* (In preparation.)

[3] H. P. Cushing, *New York State Mus. Bull. 169*, p. 19, and *Bull. 191*, p. 15, 1914.

but this may not be a safe criterion, in that quartz may have been reorganized into meta- and trisilicates.

Seeking for a more reliable distinction the pyroxene-amphibole (the pyribole of Johannsen)[1] content was examined. It is held by many geochemists[2] that pyroxene is a high-temperature mineral, while amphibole is a lower-temperature form. The change from one to the other being a paramorphic (or "autometamorphic") one —a change readily brought about by the stresses of dynamic and static metamorphism—the inversion of pyroxene to amphibole furnishes some aid in the problems in hand. If a large amount of pyroxene, such as augite, is found in an amphibolite it suggests an igneous origin. But under the stress of severe metamorphism this inversion may be complete. Martin[3] found this to be true of the amphibolite inclusions in the granitic rocks in St. Lawrence County. Thus the absence of augite does not prove a sedimentary parentage, but merely suggests it. This criterion, like the former, is therefore regarded as inconclusive.

Hunting for additional criteria, the writer investigated the feldspars in turn. It was found that the igneous types usually contained a simple range of feldspars, such as 10 per cent of soda orthoclase and 20 per cent of andesine, while the sedimentary rocks frequently exhibited a motley collection; covering a much wider range. Very commonly soda orthoclase, soda microcline, microperthite, oligoclase, and labradorite were seen in a single microscopic slide.[4]

Adams[5] states that the amphibolite occurring near Jack Lake, Ontario, to which an igneous origin must be ascribed, is "composed almost exclusively of hornblende, and plagioclase feldspar. The hornblende is rather light green in color in ordinary light. . . . . The plagioclase is clear and fresh in appearance, and rather basic

[1] Albert Johannsen, *Jour. Geol.*, XIX, p. 319, 1911.

[2] J. V. Elsden, *Principles of Chemical Geology*, p. 114, 1910; Becke, *Tsch. Mineral und Petrog. Mitt.*, *16*, pp. 327–36; F. W. Clarke, *U.S. Geol. Surv. Bull. 616*, p. 386; Lacroix, *Mineralogie de la France*, I (1893–95), pp. 668–69.

[3] J. C. Martin, *New York State Mus. Bull. 185* (1916), p. 157.

[4] H. L. Alling, *Amer. Jour. Sci.* (4), XLVIII (1919), pp. 61–62.

[5] F. D. Adams, "On the Origin of the Amphibolites of the Laurentian Area of Canada," *Jour. Geol.*, XVII (1909), pp. 1–18; F. D. Adams and A. E. Barlow, "Geology of the Halburton and Bancroft Areas," Province of Ontario, Canada, Dept. of Mines, *Geol. Survey Branch, Mem. 6* (1910), pp. 160–61.

in character . . . . [and limited to] a single species . . . . , a large proportion of [which] is frequently untwinned." The paramphibolites derived from the action of granitic intrusives and metamorphism upon Grenville limestone "are composed of quartz, microcline, orthoclase, and plagioclase."[1] This selective habit of the feldspars is explained on the ground that in the freezing of a magma the feldspars "split along the eutectic line." If the feldspar composition, in the magma, was on the potash side of the eutectic line the resulting crystals would be dominantly the orthoclase type of feldspar, while if it was on the other side plagioclase (plus a little potash feldspar) would result. But if the position of the molten feldspar was on or near the eutectic line the solid minerals would be divided on freezing into orthoclase (carrying a little soda feldspar in solid solution) and plagioclase.

Conclusion.—The criteria may be summed up as follows:

Sedimentary Origin	Igneous Origin
Original quartz	High pyroxene content
Motley collection of feldspars	Evenly "split" feldspars

How successfully these criteria have been applied to amphibolites whose origin was not forthcoming from the field relations cannot as yet be stated, but hope is entertained that some progress has been made in this difficult problem.[2]

APPENDIX[3]

THE SOLUBILITIES OF THE FELDSPAR COMPONENTS[4]

In order to understand the nature and construction of minerals from a mineralographic point of view, it is preferable to commence our consideration with reference to the state of homogeneous fusion. Although it may be that certain pairs of silicic salts cannot

[1] F. D. Adams, "On the Origin of the Amphibolites of the Laurentian Area of Canada," *Jour. Geol.*, XVII (1909), p. 10.

[2] Since this was in type a recent paper in this Journal on the "Feldspar Method" of distinguishing sedimentary and igneous metamorphics has appeared.

[3] This section is introduced to furnish the reader who is not familiar with the meaning and interpretation of thermo-equilibrium diagrams a simple explanation of their construction and value.

[4] The manner of presenting this topic has been patterned very closely after that of Rosenhain (*Introduction to Physical Metallurgy*, p. 78, 1915). It is of interest to note the similarity of mineralography to metallography.

be made to mix in all proportions while in the molten condition, it is highly probable that, in the great majority of cases, they can be mixed with one another in the fluid state in any relative proportion. In this respect these fluids resemble such liquids as water and alcohol. We may, in fact, safely carry this analogy much further, and regard mixtures of two molten silicic salts as simple solutions of the two constituents in one another. Since our interest naturally centers in the solid mineral which results from the solidification of such melts the question which lies before us is what happens to a mutual solution of two silicic salts when the temperature is lowered so that the material undergoes solidification?

The answer is that there are two opposite modes of solidification adopted by such systems, and a range of intermediate modes connecting these extremes. The one extreme is (*a*) the case in which on solidification the mineral crystallizes while still remaining a solution, i.e., the crystals which are formed ultimately attain the same composition as the molten liquid from which they crystallize. Such crystallized solutions are usually termed "solid solutions" in view of the fact that they are at the same time solids and solutions. This is the case when perfect isomorphism exists.

Such solid solutions should not be regarded as compounds, and no single chemical formula can be employed to express the mineral as a whole.

The other extreme of the mode of solidification (*b*) is that in which the state of solution which exists in the liquid condition is largely or entirely destroyed by the passage into the solid state, the two constituents separating more or less completely during the process of crystallization. This condition occurs when limited isomorphism prevails.

THE EQUILIBRIUM DIAGRAM

The most comprehensive and satisfactory method of presenting and describing the nature and constitution of minerals belonging to a given system consists in a diagram—the thermo-equilibrium diagram—which is based primarily upon thermal data. The construction of such a diagram is based upon the determination of the temperature of the specimen at various times during a heating or cooling process. The usual method consists in taking temperature

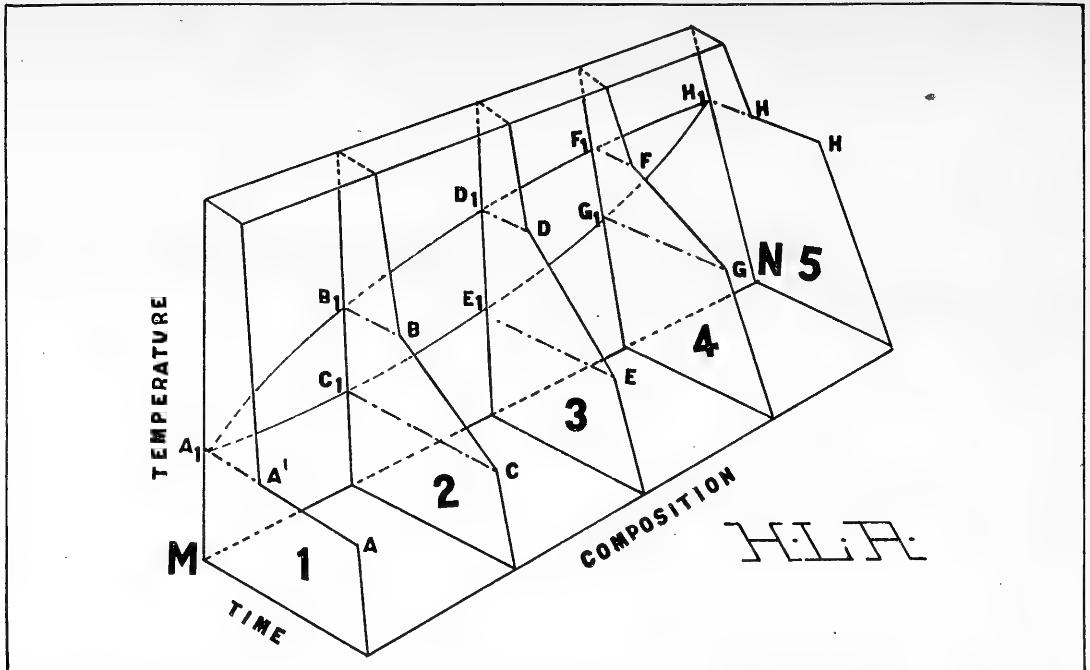


FIG. 14.—Diagram illustrating the construction of a thermo-equilibrium diagram of two components which are completely soluble in each other in the solid condition. The "partitions," 1, 2, 3, 4, and 5, are time-temperature curves, while the cigar-shaped area is the conventional form of the equilibrium diagram.

readings at fixed intervals of time and then plotting the results with temperatures as ordinates and times as abscissas. A time-temperature curve is thus obtained which indicates the behavior of the mineral in a most direct way. So long as the substance is simply raised or lowered in temperature at a steady rate, this curve follows a smooth course; a departure from this smoothness indicates that there has been either an evolution or an absorption of heat within the specimen. Such a change in the shape of the curve indicates a change of state, either in phase, or in modification. Time-temperature curves are obtained for a binary system from specimens composed of the two constituents in varying amounts from 100 per cent of one to 100 per cent of the other. The construction of the equilibrium diagram from these time-temperature curves is illustrated in Figure 14. Five time-temperature curves are shown as partitions in the end of a box, numbered 1, 2, 3, 4, and 5. The critical points or places where the curves change in direction are indicated by A', B, D, F, and H'. These mark the points where crystallization commences and A, C, G, and H, indicate where the solidification is complete, if the specimen is allowed to cool. If, however, the specimen is reheated, it will theoretically at least pass through the identical behavior except in the reverse order. That is, the points A, C, G, and H are determined by the initial melting, and the points $A'B$, D, F, and H' by complete liquefaction.

Now these time-temperature curves (partitions) enable us to construct the equilibrium diagram by projecting or drawing construction lines parallel with the base from the points already mentioned back to the vertical plane, as is indicated in Figure 14. Removing the time-temperature curves, which are merely scaffolding, the diagram remains as a conventional method of indicating the crystallization behavior of an isomorphous series, i.e., a series of solid solutions.

In Figure 15 the opposite extreme of a binary system is shown. The diagram is constructed in the same manner, from the projection of the critical points of a series of time-temperature curves. In this case the two components are completely insoluble in the solid state. It will be noticed that the upper line, $A_1B_1E_1$, repre-

senting the freezing temperatures (and called the liquidus) falls off or is depressed as amounts of the second component are added,

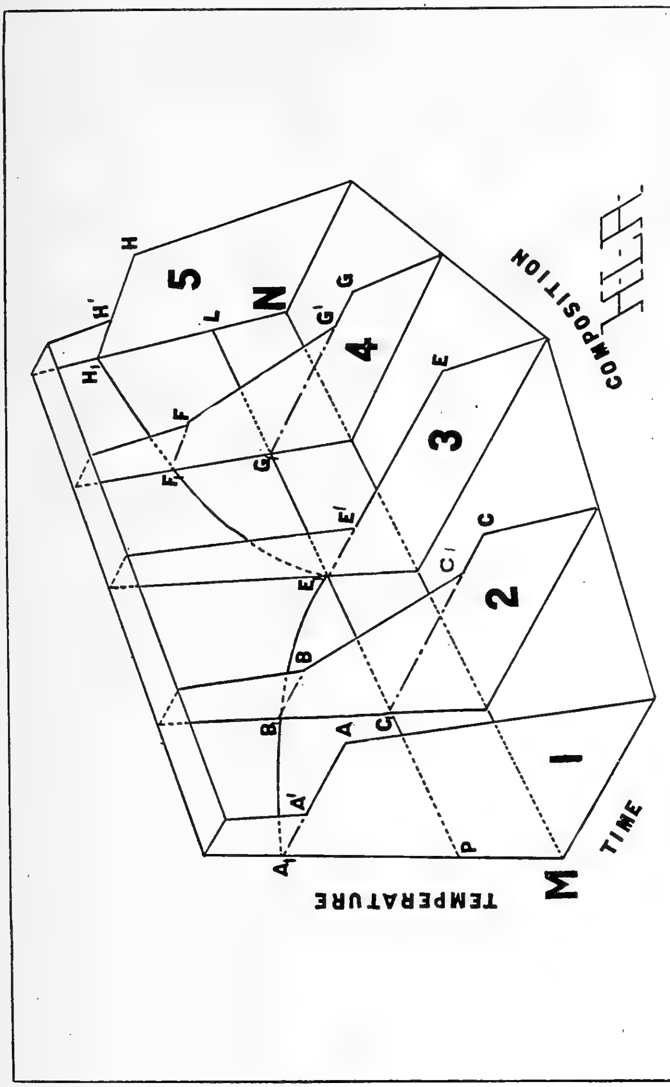


FIG. 15.—Diagram illustrating the construction of the thermo-equilibrium diagram of a eutectiferous system

until (near the center of the present diagram) a point is reached where the sum of the two components freezes at a temperature lower than in any other proportion. This is the "eutectic point"

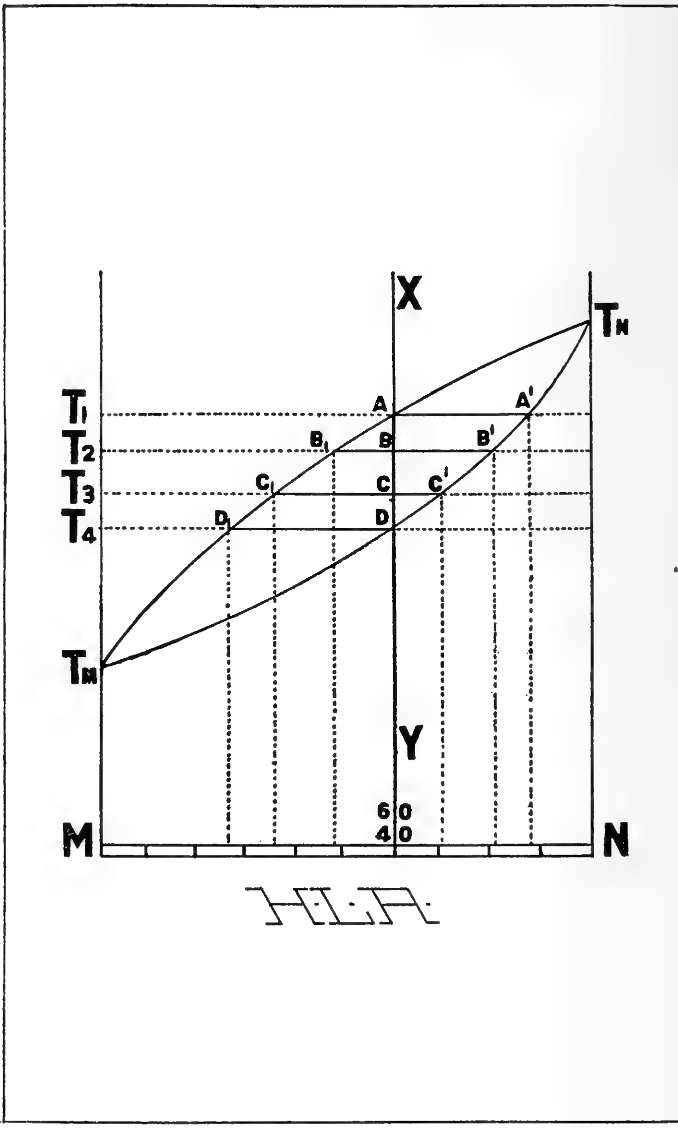


FIG. 16.—Thermo-equilibrium diagram of a series of solid solutions. Composition of melt taken as 60 per cent of N and 40 per cent of M. For discussion see text. Roozeboom's Type I.

(εὖ, easy, τηκτός, melting). The line $PC_1E_1F_1L$ is the other curve of the diagram, the solidus.

Now having described the manner in which an equilibrium diagram is constructed let us see what it can tell us of the process of solidification of an isomorphous series. Figure 16 is a diagram of such a system (Type I of Roozeboom's classification), the time-temperature construction curves not being shown. The temperatures form the ordinate and the composition the abscissa. Above the liquidus line $T_MD_1C_1B_1AT_N$ the solution of the two components as a liquid is a mutual one, that is they mix in all proportions. Between the liquidus and the solidus—the lens-shaped area—is the area in which solidification is going on and is occupied by both liquid and solid phases. Below the solidus the system is solid. Now let us trace in detail what happens during the freezing of a melt composed of 60 per cent of *N* and 40 per cent of *M* as is indicated by the vertical line *XY*. Above *A* the system is liquid, but as the temperature, falling, reaches the liquidus line, crystallization commences, precipitating crystals of the composition A'. The composition of such a crystal is obtained by constructing the horizontal dash line from *A* to the solidus. Thus the first crystal formed has a composition of A'. The remaining liquid has a composition of *A*. With continued lowering of the temperature the point represented by *B* is reached. The composition of the crystal is B' and that of the liquid is B_1.

Temperature	Composition of Crystal	Composition of Liquid
T_1	A'	A
T_2	B'	B_1
T_3	C'	C_1
T_4	D	All liquid frozen

It will readily be seen that in the phenomenon above, where we are assuming, for the time being, that no adjustment of the crystals between themselves or between the crystals and the liquid takes place, the resulting crystals will have a wide range in composition, or a single crystal will be built up concentrically of zones of variable composition. It follows from the examination of the diagram that the center of such crystals will have a

composition richer in the component that possesses the higher freezing-point than the margins. Such zonal crystals are common among the plagioclase feldspars. Not all zonal textures, however, are to be explained in this way, as will be explained later in more detail. All available information leads to the conclusion that this type of diagram is the one to which the plagioclase feldspars (the soda-lime series) are to be assigned. In the field of mineralography our knowledge of other and similar systems is very incomplete. In addition to the plagioclase feldspars, the garnets, the scapolites, the micas, the alums, and certain ranges of the pyroxene and amphibole families may be mentioned. But in the field of metallography, binary systems of solid solutions are better understood. The following systems may be noted: Ag-Au, Ag-Pd, Au-Pd, Bi-Sb, Co-Fe, Co-Ni, Cu-Pd, Cu-Pt, Fe-Mn, In-Pb.[1] There is no fundamental difference between isomorphous minerals and the alloys belonging to this group.

Now let us examine in more detail the other extreme (*b*), that of a eutectiferous system, the two components of which are entirely insoluble in each other when in the solid state. In Figure 17 the composition of the melt chosen is the same as that used before, namely 60 per cent of N and 40 per cent of M. On cooling the melt the freezing commences at A; the resulting crystal having a composition of A_1 or pure N, the liquid, a composition of A. This can be expressed by saying that the composition of the crystal "varies" or "slides" down the line $A_1B_1C_1D_1$ while that of the remaining liquid "slides" down the liquidus from A through B and C toward E where it freezes, but on passing to the solid phase the solution E is rendered a mechanical mixture as the two components separate, theoretically at least, into pure N and pure M. This mechanical mixture is termed the eutectic mixture. This tells us how such a system will look under the microscope, either in thin sections or as etched polished slabs. It will consist of crystals of pure N surrounded by minute crystals of pure M and pure N. The groundmass of the small crystals will be the eutectic mixture. It is possible to reverse the scheme and determine the composition of the original melt by determining the

[1] C. H. Desch, *Metallography* (1913), p. 401.

amount of the crystals of pure N and of the small crystals which constitute the eutectic mixture, and "sliding" back up the lines until A is reached.

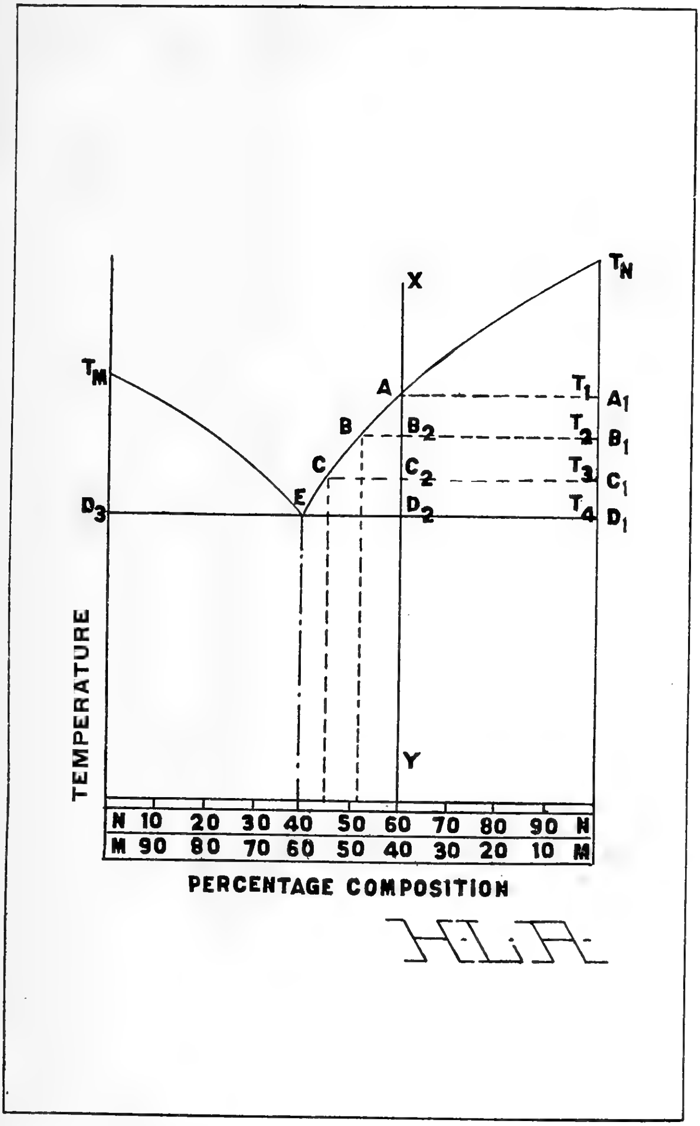


FIG. 17.—Thermo-equilibrium diagram of a eutectiferous system

We are not greatly interested in Figure 17 for it was introduced as a stepping-stone to Figure 18 with which we shall have much to do. Intermediate between the two diagrams already discussed is a type of diagram shown in Figure 18 (Roozeboom's Type V) which represents a condition where limited solubility between the two components prevails in the solid condition. The lines FD_3 and GD_1 are the assumed solubility lines, showing the limits of solubility. Again, going through the same procedure as before, we take a melt composed of 60 per cent of N and 40 per cent of M and allow the temperature to fall to T_1 indicated by the point A on the liquidus. At this point the composition of the crystals separating into solid form is A. This is not pure N as it would be in the previous Figure 17 but it is a solid solution, 91 per cent N and 9 per cent M, the percentage being found by dropping a perpendicular line from A_1 to the "Percentage Composition" scale at the bottom of the Figure 18. The liquid remaining has an approximate composition of A. As the temperature falls the change in the composition of the crystals is represented by the liquidus curve from A through B and C toward E. When the eutectic point E is reached the eutectic mixture freezes, being composed of solid solution D_1, 80 per cent N and 20 per cent M, and solid solution D_3, 20 per cent N and 80 per cent M. Thus the resulting solid mineral is composed of crystals—solid solutions—having a range in composition represented by the curve A_1D_1 surrounded by the eutectic mixture which is composed of two solid solutions, D_3 and D_3.

The discussion of these diagrams has been merely for the purpose of attempting to explain the meaning and the use of what are known as thermo-equilibrium diagrams.

CONCLUSIONS

1. The application of the phase rule and thermo-equilibrium diagrams to the feldspar system enables the mineralogist and the petrographer to secure a much better conception of the true physical-chemical nature of these minerals. This method of investigation throws considerable light upon the character of many other mineralogical systems.

2. The thermo-equilibrium diagram of the iron-carbon binary system offers the mineralogist a method of studying the alloys which

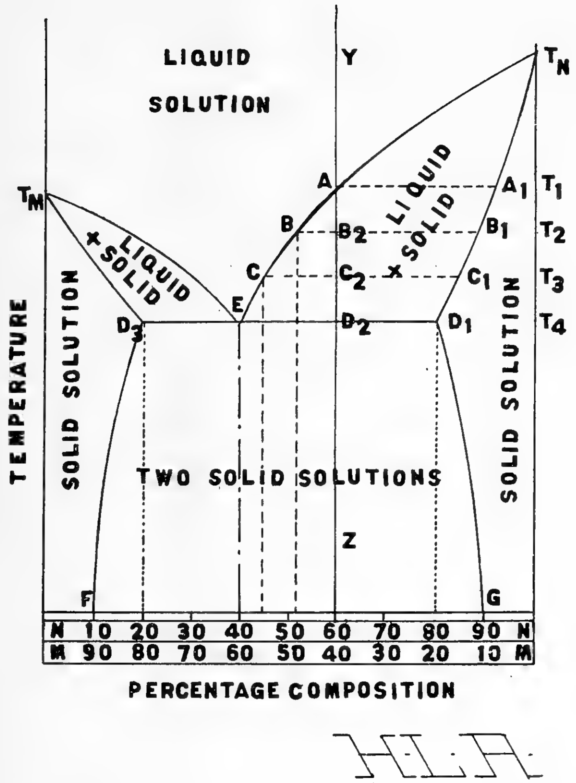


FIG. 18.—Thermo-equilibrium diagram of a eutectiferous, limited-solubility system. The general type probably represented by the potash-soda series.

in his domain are called minerals. Familiarity with this system is of great assistance in clarifying the problems of mineralogy.

3. The feldspars belong to a five-component system, of which the end members are the potash, soda, lime, barium, and carnegieite feldspars. For most purposes it is only necessary to consider the potash, soda, and lime components. Every specimen of feldspar found in nature contains a certain amount of each of these three components.

4. The plagioclase and the hyalophane series constitute a series of solid solutions. The potash-soda and the potash-lime series possess only limited solubility, and constitute eutectiferous systems.

5. It is believed that both the potash and the soda feldspars are dimorphous, each existing in two isomeric forms: each component crystallizing either in monoclinic or triclinic modifications, depending upon the temperature and the viscosity of the magma; that orthoclase and albite are high-temperature modifications and that microcline and possibly (?) barbierite are relatively low-temperature forms.

6. The complete solubility of albite in anorthite, and vice versa, presupposes that their chemical structures are analogous, even though albite is usually regarded as a "trisilicate" and anorthite an "orthosilicate." This presupposes that the feldspars are aluminosilicates. Since orthoclase and albite (components) are not completely soluble in each other (when perfect equilibrium prevails), they probably possess somewhat dissimilar chemical structures.

7. Some feldspars contain nephelite in solid solution but this mineral cannot be regarded as isomorphous with the normal feldspars. Therefore the mineralographic term "solid solution" is more comprehensive than the crystallographic term "isomorphism."

8. All feldspars are solid solutions and mixtures of solid solutions, and therefore do not possess definite chemical compositions. No single chemical formula can be assigned to a single species. Inasmuch as labradorite is a mineral, the chemical composition of which is *not* fixed, and, furthermore, the mineral is often found zonally grown, the usual definition of a mineral as a *homogeneous*

natural inorganic substance of *definite chemical composition* needs revision.

9. That the plagioclase feldspars are not "molecules," portions of which are "replaceable" by analogous units. This statement applies with almost equal force to the potash-soda series as well.

10. While the term "mixed crystals" is frequently used to signify "solid solutions," yet it should be avoided for the sake of clearness.

11. Perthite is an intergrowth of two (or more) solid solutions, and not an intergrowth of the simple components, microcline (or sometimes orthoclase) and acid plagioclase. The two phases are solid solutions, one rich in potash and the other rich in soda.

12. That most perthites are not the *direct* result of the freezing of a magma, but are the result of *subsequent* processes, where the decrease in solubility of one phase for the other with falling temperature is the principal factor. The inversion of orthoclase to microcline is regarded by some as also a contributing cause. Perthites are commonly the result of the process here called "exsolution." Such perthites (or "perthoids") are analogous to pearlite in steels.

13. That many anorthoclase specimens are supersaturated, undercooled metastable solid solutions, potentially perthite through the intermediate stages of crypto- and microperthite.

14. That intergrowths of potash-rich and lime-rich solid solutions occasionally are found. To such intergrowths the term *oranite* (the first two letters of *or*thoclase and *an*orthite and the ending *-ite*) has been applied.

15. That the feldspars of many basic monzonites and the granodiorites are approaching, as the limit, the potash-lime binary system. But because such feldspars are not as viscous at their melting temperatures as the potash-soda series, they separate more completely into definite identities, and consequently the *oranitic* feldspars are not usually recognized as such.

16. The significance of the process of exsolution is that many so-called inclusions in mineral grains are due to secondary processes, and consequently are of late rather than of early development.

This complicates the methods of determining the order of crystallization of the minerals in a rock.

17. Primary perthite, due to the freezing of the eutectic mixture, analogous to ledeburite in steels, is probably uncommon in nature.

18. The potash feldspar of pegmatitic origin, the usual museum variety of "orthoclase," is soda microcline and very rarely, if at all, "orthoclase." In fact orthoclase (nearly pure potash feldspar without microclinic characteristics) is very rare in nature.

19. Some "adularias" and "microclines" show microclinic twinning when examined in thin sections, but in thin plates or in crushed fragments they do not exhibit twin striations. The suggestion is strong that the pressure, which the specimens experienced in the grinding process in the preparation of the thin section, has hastened the inversion of the metastable soda orthoclase to soda microcline. This raises the question whether complete reliance can be placed upon thin sections in the identification of the feldspar species.

20. That microclinic and orthoclasic feldspars, with a content of the potash component higher than 85 per cent of the whole, are exceedingly rare in nature. It is far more accurate to speak of *soda* microcline and *soda* orthoclase than of microcline and orthoclase. Very frequently specimens of so-called microcline or orthoclase are found upon examination to be perthitic as well. Before assigning a name to a museum specimen, it should be microscopically examined.

21. That the inversion of soda orthoclase to soda microcline is often hastened by the pressure set up by regional or static metamorphism, but that the pressure does not *produce* soda microcline from soda orthoclase; it only initiates and accelerates the change. The tendency to change is inherent; the pressure merely starts the process.

22. All plagioclase specimens contain some potash component; the average is in the neighborhood of 5 per cent. It is more accurate to assume that the potash component is present to this extent than it is to ignore it altogether. The extinction angles of the soda-lime feldspars enable the petrographer to ascertain the amount of the soda component present but they do not determine the amount of the lime component. The percentage of the potash

feldspar is inferred from the average of the chemical analyses and the lime component is the remainder of the 100 per cent.

23. Many zoned plagioclase feldspars are to be explained by the process of normal crystallization under rapid chill instead of "magmatic corrosion." Homogeneous crystals of plagioclase are probably due to readjustment between crystal phases, or between them and the surrounding unfrozen liquid, in a slowly cooling magma. The degree of homogeneity is therefore a function of the rate of chill. Some zoned feldspars are undoubtedly the result of more complex processes in which the phenomenon of undercooling plays an important rôle.

24. The physical properties of a series of solid solutions are direct functions of the composition. If the properties, such as specific gravity, indices of refraction, extinction angles, etc., are plotted in conjunction with the thermo-equilibrium diagram they assume the form of smooth curves which rise and fall with the freezing (liquidus) curve. A break, a cusp, or a sharp change in direction in these curves at least suggests a discontinuity in the chemical properties of the system. Many involved formulas of minerals will probably be abandoned when they are shown to be solid solutions and mixtures of solid solutions of simple end members. The formula for hyalophane, "$(K_2,Ba)Al_2(SiO_3)_4$," cannot be entertained as possessing any true value.

25. In attempting to classify the feldspars on a basis of their true composition several new names have been proposed. See Figure 13 for schemes of classification. "Anorthoclase" is defined in paragraph 13 and its range delimited as $Or_{70}Ab_{30}$-$Or_{35}Ab_{65}$. The term "soda microcline" is confined to microclinic feldspars containing 10 to 30 per cent of the soda component. For albitic feldspars containing from 5 to 20 per cent of the potash component the term "potash albite" is proposed.

26. Because there are more contours of equal extinctions (isogonic lines) cutting the plagioclase side of the triangle (see Fig. 12, facing p. 251) than the number of those cutting the potash-soda side, the identification of the soda-lime feldspars can be accomplished with comparative ease. Nevertheless proper identification of the subspecies of the potash-soda series can be

made by means of the extinction angles on the different faces, together with the indices of refraction.

27. The application of these physical-chemical principles to the feldspar system furnishes, as a result, the means of solving practical geological problems. Two illustrations of the application of these conclusions to actual field problems are given on pages 275 to 279 inclusive.

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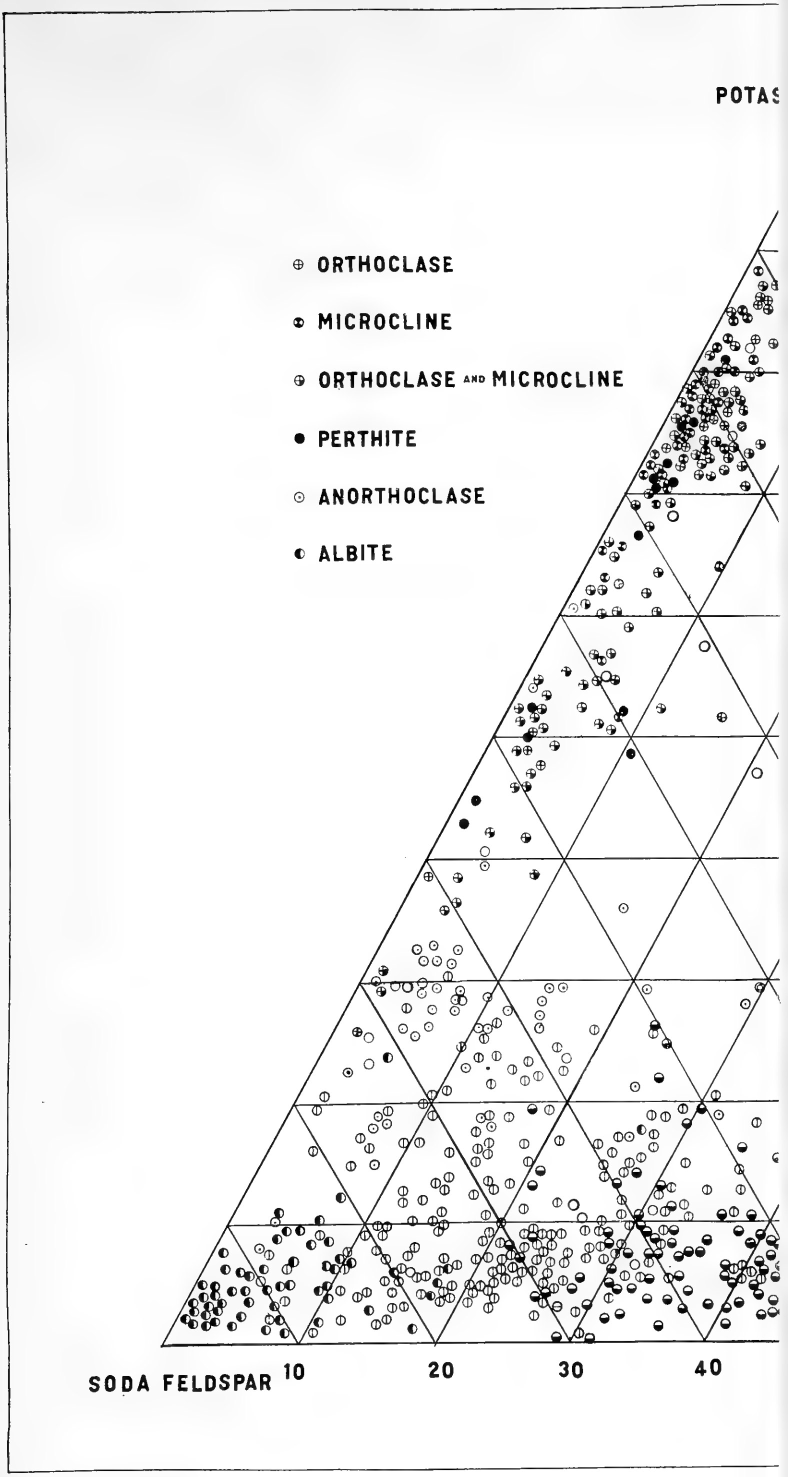

FIG. 19.—Plot of 954 recast chemical analyses of natural feldspars. The chemical analyses and are not those which the writer would employ in many cases plagioclase feldspars contain considerable amounts of the potash component. The

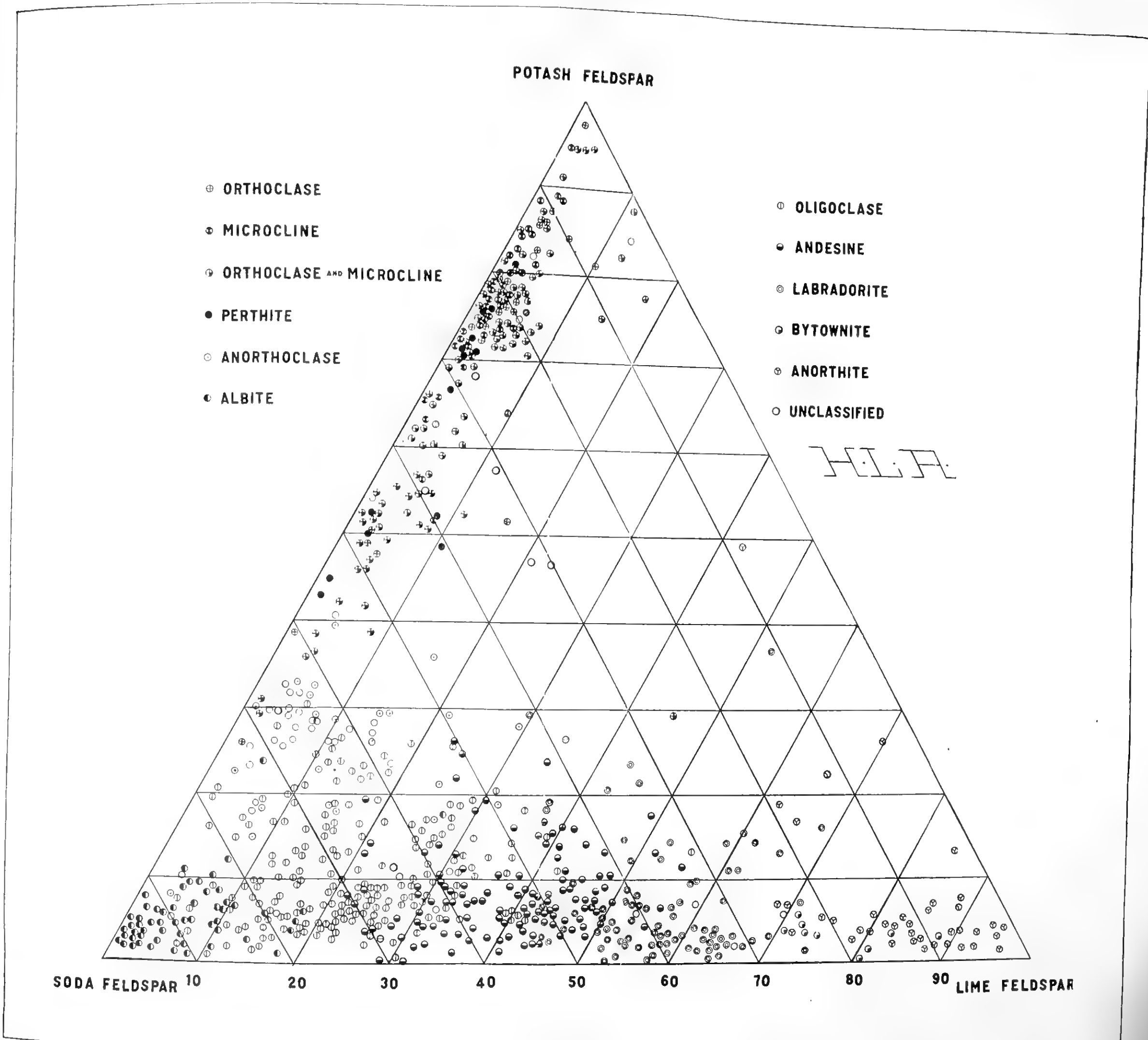

FIG. 19.—Plot of 954 recast chemical analyses of natural feldspars. The names assigned to the species here represented are those published in connection with the chemical analyses and are not those which the writer would employ in many cases. Note the concentration of "microcline" and "orthoclase" specimens and the fact that most plagioclase feldspars contain considerable amounts of the potash component. The almost total absence of representatives of the potash-lime feldspars is also shown.

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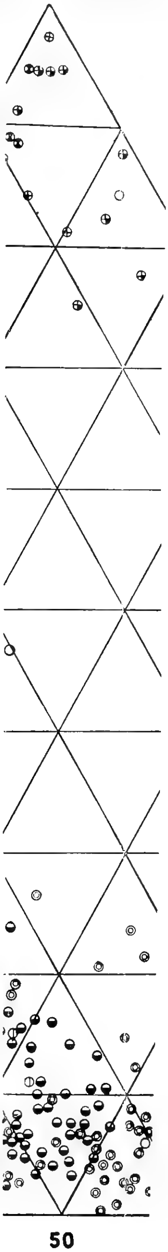

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DIFFUSION IN SILICATE MELTS

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INTRODUCTION

Many well-known features of igneous rocks are clearly the result of diffusion. The growth of crystals is accomplished largely by the diffusion of material to crystalline nuclei. Inclusions often exhibit solution at their borders and some diffusion of their material into the surrounding magma. These phenomena involve diffusion through short distances and there is no theoretical objection to the acceptance of diffusion at the required rate. At the same time it has often been inferred that large quantities of matter have diffused through considerable distances. This has been done for the purpose of explaining various associations of igneous rocks and especially the formation of basic border phases. Since Becker's destructive criticism of this view[1] many geologists, but not all, have been less willing to assign to diffusion any important rôle in the production of such features.

Becker's objection was based on the extreme slowness of diffusion in all cases where its rate had been determined. The principal data then available referred to the diffusion of salts in aqueous solution, and it was on the basis of reasonable deduction

[1] *American Journal of Science* (4) Vol. III (1897), p. 27.

from these data that Becker's objection was raised. There has been as yet no discovery of a general principle connecting the rate of diffusion of matter in solution with other physical constants. Each substance has its coefficient of diffusivity (in a given medium) characteristic of the substance and determinable only by experiment. The coefficient is the constant factor, k, in the equation expressing Fick's law of the rate of diffusion, viz.,

$$\frac{dc}{dt}=k\cdot\frac{d^2c}{dx^2},$$

and is evidently equal to the number of grams which diffuse past 1 square centimeter of any plane in unit time when the concentration gradient normal to the plane is unity. This is a very small quantity for substances investigated. For most salts in aqueous solution it is of the order 3×10^{-6} in cm.^{2} per second at the ordinary temperature and increases rapidly with rise of temperature; for molten metals it is considerably larger and of the order 3×10^{-5}, again increasing with rise of temperature.[1]

In the case of the quite different type of matter, molten silicates, it is perhaps not probable, but nevertheless not inconceivable, that the coefficient of diffusivity should be fairly large at the high temperatures concerned in spite of the usual high viscosity. Only actual measurements can make us at all certain that we have a firm basis of fact for the discussion of diffusion in molten silicates. Practically no measurements have been made. Endell has demonstrated the fact of the interdiffusion of lime and microcline glass.[2] Schulze has measured the rate of migration of silver ion in glass.[3] Indeed, it is perhaps principally from the problems of glass manufacture that we gain our impressions of diffusion in molten silicates. There diffusion is often excessively slow, but it should be noted that glasses belong for the most part among the very viscous silicate mixtures. On account of the lack of experimental data it was considered advisable to undertake some measurements of the

[1] Roberts-Austin found for the coefficient of diffusivity of gold into molten lead $k=3.47\times10^{-5}$ at 492°. *Roy. Soc. London, Phil. Trans.*, Vol. 187A (1896), p. 383. See also Van Orstrand and Dewey, *U.S. Geol. Survey, Prof. Paper 95-G.* 1915.

[2] K. Endell, *Silikat-Zeitschrift*, Vol. I, p. 195.

[3] *Ann. phys.*, Vol. XL (1913), p. 335.

rate of diffusion in fused rock-forming silicates. Those here described are to be considered of a preliminary nature. They are not devised with the purpose of establishing precise values for diffusion coefficients from which general theoretical conclusions might be drawn, though this is recognized as a desirable ultimate goal. They may serve rather to aid the geologist in deciding what he may and what he may not reasonably attribute to diffusion.

THE METHOD OF EXPERIMENT

The method followed was that of permitting the diffusion against gravity of a heavy liquid placed in the lower part of a crucible into a lighter liquid in the upper part. Ostwald has said that to make accurate experiments on diffusion is one of the hardest problems in practical physics on account of the difficulty experienced in eliminating convection currents. It is to be noted that this is especially true of aqueous solutions. Water is a thin liquid with a relatively high coefficient of thermal expansion. The driving force of convection, viz., difference of density, is therefore large and the resistance to it (viscosity) small. In silicates, however, these conditions are reversed, the viscosity being relatively great and the thermal expansion relatively small. While a small difference of temperature may establish convection in aqueous solutions, it is not to be expected to have a comparable effect in silicates. In the case of aqueous solutions the density gradient resulting from the composition gradient may be very small, but in the case of the silicates used it is quite large. If the relative densities of diopside and plagioclase liquids at high temperatures are comparable with those at lower temperatures, it can readily be shown that it would be necessary to have a temperature gradient of 20° per mm. in order to counteract the density gradient due to 1 per cent per mm. change of composition. Gradients of composition of this magnitude exist throughout most of the period of experiment. Such temperature gradients are, however, entirely lacking; not only this, but it is easy to make the moderate temperature gradient that does exist of such a sign that it acts together with the composition gradient instead of counteracting it. This is accomplished, of course, by making the temperature of the upper

part of the column higher than that of the lower part. With this end in view the crucible was suspended in most cases in a part of the furnace where the temperature increased upward. In one experiment recorded here this method of eliminating convection was replaced by a method involving the use of a bath of molten gold to obtain a uniform temperature. The platinum crucible containing the charge was protected from the molten gold by a tube of silica glass.[1] The silica glass was rather soft at the temperature of the experiments and required reinforcement by a tube of Marquardt porcelain. The one result obtained by this method showed no significant difference from a result obtained by the method of suspending the bare crucible in the furnace. There seemed, therefore, to be no reason for preferring the use of the gold bath and it was not carried farther.

To make assurance doubly sure in the way of minimizing possible convection, the charge was made small. This acts in three ways: to make a possible lateral difference of temperature small, to render difficult the initiation of convection currents, and to decrease the velocity of possible currents. The crucibles used were, therefore, 5 to 6 mm. in diameter and 10 to 20 mm. deep. In order to simplify the conditions of diffusion the crucibles were right circular cylinders without flare or rounded bottoms.

The temperature was kept constant partly by continual watching and regulation and partly by using current from a storage battery. In some cases an automatic regulator was used, designed by W. P. White of this Laboratory.

In each case the heavy material taken was diopside. It was first melted and then chilled to a firm cake of glass $\left(G\frac{20}{4}=2.854\right)$ in the bottom of the crucible. The lighter material put in the top was one of the plagioclases Ab_2An_1, Ab_1,An_1, or Ab_1An_2 $\left(G\frac{20}{4}=\right.$ $\left.2.483,\ 2.533,\ 2.591\right.$ respectively). The temperature was in all cases about 1500°, that is, it was above the melting temperature of both layers, so that the experiments deal with diffusion of one liquid silicate into another.

[1] The use of silica glass in this manner was suggested by J. C. Hostetter, formerly of this Laboratory, who also kindly worked the glass into the desired form.

The charge was raised quickly to the desired temperature by plunging it into a furnace already somewhat above that temperature. After holding it for the desired length of time the charge was rapidly cooled by removal from the furnace, and the composition of the glass at various levels in the crucible was determined by measuring its refractive index, the relation between composition and refractive index having been previously determined on mixtures of known composition.

It may appear that the drastic temperature differences that arise when the cold crucible is placed in the hot furnace would be bound to set up violent convection, but fortunately this is a matter that can easily be ascertained by running a blank test. For this purpose a charge was prepared in the ordinary manner and allowed to remain in the furnace only a few minutes, when it was removed and the distribution of composition determined. It was found that the distribution was as it should be after diffusion for a short period with no random variations such as would result from convection.

In earlier experiments an ordinary thick-walled platinum crucible was used and the charge was always badly shattered in cooling. When not too numerous, the fragments were fitted together to reconstruct the original charge and the composition determined at various levels by removing a little powder with a file and determining its refractive index by the immersion method. The error involved in the measurement of the distance from the top or the bottom of the charge was large, and the results were only rough approximations. In later experiments the crucible was made of platinum foil 0.03 mm. thick. On contracting, the glass pulls the weak walls of the crucible with it and remains unshattered. The platinum foil was then peeled off and the cylinder of glass was ground to a wedge whose edge was parallel to the axis of the cylinder. The faces of the wedge were polished and the refractive index of the glass at various points was determined on the goniometer by the method of minimum deviation. The exact distance of the points from the bottom of the cylinder was measured by means of the scale on the centering screws of the goniometer.

Though numerous measurements were made by the early, rougher method, only the few made by the later method will be described. It may be noted, however, that the earlier results, within their rather large limits of error, agree with the later.

In order to obtain a visual impression of the diffusion of material, such as is obtained when copper sulphate crystals are placed in the bottom of a vessel of water, the layer of diopside was in some cases colored by the addition of 1 per cent Fe_2O_3 which imparts to the glass a marked green color. In all cases the change of color was sensibly coincident with the change of refractive index; in a layer where the index fell rapidly the color faded rapidly; in cases where an upper layer was found unaffected in refraction it was likewise uncolored.[1]

In summary it may be stated that the mode of procedure was as follows: A layer of diopside was placed in the bottom of a crucible with a layer of plagioclase above it, diffusion was permitted at constant temperature (above the melting-temperature of both layers) for a definite period, the charge was quenched, and the composition determined at various depths by measuring the refractive index of the glass.

GRAPHICAL PRESENTATION OF RESULTS

The results (Table I) may be presented graphically as in

TABLE I

Diffusion Results

EXPERIMENT NO. 24

Depth of diopside layer 1.8 mm. Total depth 9.5 mm. Time 48 hrs.

Distance from bottom (mm.)	0.3	1.3	2.3	3.3	4.3	5.3	6.3	7.3	8.3	9.3
Vol. per cent diopside	39.5	39.6	38.1	31.5	18.0	7.5	4.8	3.5	1.1	0

Plagioclase of upper layer Ab_2An_1.

EXPERIMENT NO. 27

Depth of diopside layer 3.2 mm. Total depth 11.7 mm. Time 48 hrs.

Distance from bottom (mm.)	1.0	2.0	3.0	4.0	5.0	6.0	7.0	8.0	9.0	10.0	11.0
Vol. per cent diopside	52.0	51.6	50.7	49.6	46.5	33.7	9.7	2.4	0.7	0	0

Plagioclase of upper layer Ab_2An_1.

[1] This result might be taken as indicating that selective diffusion was unimportant, though not necessarily absolutely lacking.

EXPERIMENT NO. 21

Depth of diopside layer 3.9 mm. Total depth 9.1 mm. Time 22 hrs.

Distance from bottom (mm.)	0.2	1.0	2.0	4.0	5.0	6.0	7.0	8.0	8.8
Vol. per cent diopside	48.0	47.9	47.5	45.5	43.0	41.0	36.5	30.5	29.5

Plagioclase of upper layer Ab_1An_1.

Immersed in bath of molten gold.

EXPERIMENT NO. 37

Depth of diopside layer 7 mm. Total depth 10.2 mm. Time 17 hrs.

Distance from bottom (mm.)	0.3	1.3	3.3	4.3	5.3	6.3	7.3	8.3	9.3	9.5
Vol. per cent diopside	73.0	72.8	72.3	70.5	70.7	68.5	66.5	64.2	63.6	63.4

Plagioclase of upper layer Ab_1An_1.

EXPERIMENT NO. 25

Depth of diopside layer 3.2 mm. Total depth 14.5 mm. Time 22.5 hrs.

Distance from bottom (mm.)	1.3	2.3	3.3	4.3	5.3	6.3	7.3	8.3	9.3	10.3	11.3	12.3	13.3	14.2
Vol. per cent diopside	42.5	38.3	36.3	29.8	24.4	21.3	20.1	17.8	17.2	15.0	14.6	9.5	9.0	7.3

Plagioclase of upper layer Ab_1An_2.

Figures 1–5, in which the ordinates represent height in millimeters above the bottom and the abscissae composition in units per cent of diopside. The initial condition will then be represented by two vertical lines indicating two uniform layers, one at 100 per cent diopside and of a length corresponding to the depth of the diopside layer, the other at 0 per cent diopside representing the layer of plagioclase above it and of appropriate length. These are joined by a horizontal line indicating instantaneous change of composition. The final condition will be represented by a curve on which any point represents the composition at the corresponding level. The slope at any point indicates the composition gradient at that point, the curve approaching more nearly to the horizontal the greater the composition gradient. The figures need little discussion since they present the results better than can be done in words.

THEORETICAL CONSIDERATIONS

As noted on an earlier page, concentrations in a mass undergoing diffusion may be calculated on the basis of Fick's law. In its application to the present case, and in the form most useful for

calculation of the concentration at any point in a diffusion cylinder, the equation[1] becomes

$$2c = c_0 \left[\frac{2}{\sqrt{\pi}} \int_{\frac{-l-x}{2\sqrt{kt}}}^{\frac{l-x}{2\sqrt{kt}}} e^{-\beta^2} d\beta + \frac{2}{\sqrt{\pi}} \int_{\frac{-l-(2m-x)}{2\sqrt{kt}}}^{\frac{l-(2m-x)}{2\sqrt{kt}}} e^{-\beta^2} d\beta + \text{etc.} \right] \quad \text{I}$$

where the term in brackets in the limits is successively x, $2m-x$, $2m+x$, $4m-x$, $4m+x$, etc., and where c is the volume concentration at any point at distance x from the base of the column, l is the thickness of the bottom layer of original uniform concentration c_0, m is the total length of column, t is the time elapsed, and k the constant of diffusivity. For the examples in hand the series is rapidly convergent. With the aid of this equation we may, then, calculate the concentration at various points after a certain period of time and for a certain value of k (or, more simply, for a certain value of the product kt) and draw a curve representing the theoretical distribution of concentration. Curves of this kind were drawn and it was found that in no case could a calculated curve be obtained that would coincide with the observed curve. Of the calculated curves a certain one was chosen and was plotted on each of the figures as a dotted curve. The theoretical curve chosen in each case was that which showed approximately the same concentration at the upper surface as that actually found. The curves therefore coincide at their upper ends, but at other depths wide divergence is shown between the full curve of actual concentration and the dotted curve of theoretical concentration. In all cases this divergence is of a systematic kind, the actual concentration showing a smaller gradient in the diopside-rich layers and a larger gradient in the diopside-poor layers than the theoretical concentration. This is shown particularly plainly in Figure 1, where the diopside-rich layers have reached practical uniformity while the upper layers show a very strong gradient. This uniform

[1] The equation is not so formidable as it appears, $\frac{2}{\sqrt{\pi}}\int_0^q e^{-\beta^2} d\beta$ being merely the probability integral whose value, for various values of q in the limits, can be looked up in the tables.

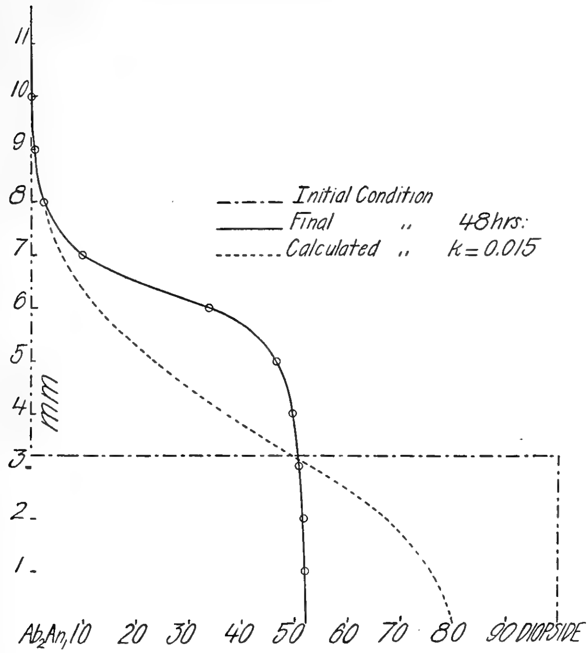

FIG. 1.—Diffusion experiment No. 27

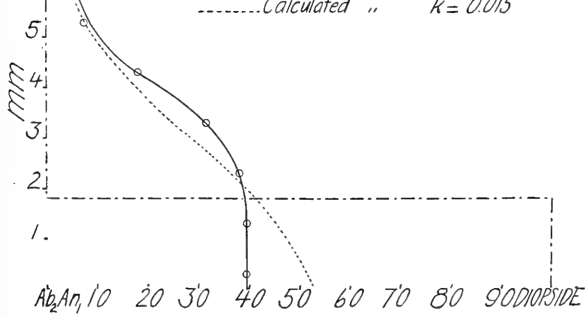

FIG. 2.—Diffusion experiment No. 24

divergence from theory may be explained by assuming that the coefficient of diffusivity is not a constant but is itself a function of concentration and is greater for diopside-rich mixtures than for those poor in diopside. In those experiments with aqueous solutions, where the highest degree of correspondence with theory is obtained, the solutions are always kept dilute so that the medium into which diffusion is taking place is sensibly constant. Under these conditions theoretical concentrations calculated on the basis of a constant value of the coefficient of diffusivity are in marked accord with observed values. In the present case, however, there is a continual and very important change in the nature of the diffusion medium as time progresses, and no constancy is to be expected in the value of the coefficient of diffusivity. There is no necessity, therefore, for regarding the results as showing divergence from Fick's law, the results being reconcilable with theory if it is assumed, as mentioned above, that the diffusivity is a function of concentration.

Einstein has developed for dilute solutions a theoretical relation between diffusivity and viscosity which makes them inversely proportional.[1] While experimental results do not entirely confirm his theory, they suggest its correctness if certain disturbing factors such as hydration could be evaluated.[2] At any rate, if we assume that the diffusivity is an inverse function of viscosity, we obtain a natural explanation of our experimental results. The viscosity of the diopside-rich mixtures is much less than that of the plagioclase-rich mixtures. The coefficient of diffusivity for the diopside-rich mixtures should be correspondingly greater, and this we have found to afford a natural explanation of the deviation from theoretical values calculated in the ordinary way. Moreover, as one would expect, this deviation is more marked for the plagioclase liquid Ab_2An_1 whose viscosity is very much greater than that of diopside liquid, and less marked for the plagioclase liquid Ab_1An_2 where the viscosity contrast is not so great, while liquid Ab_1An_1 occupies an intermediate position in this particular (cf. Figs. 1, 3, and 5). Qualitatively, then, the experimental results suggest

[1] *Ann. d. Physik*, Vol. XVII (1905), p. 549.

[2] L. W. Öholm, *Medd. K. Wetenskapsakad Nobelinstitut* 2, No. 26, p. 21.

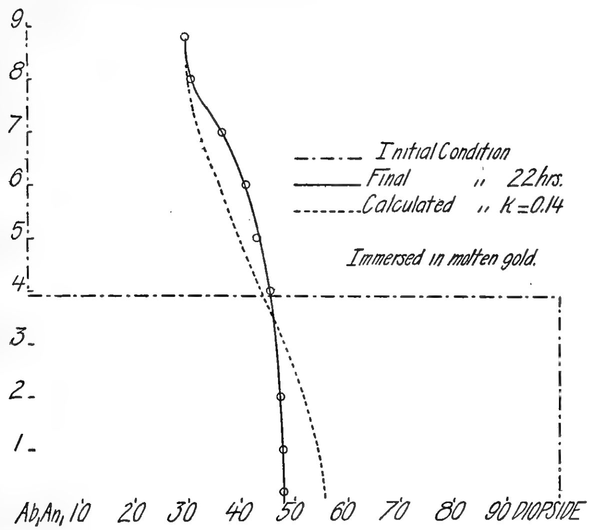

FIG. 3.—Diffusion experiment No. 21

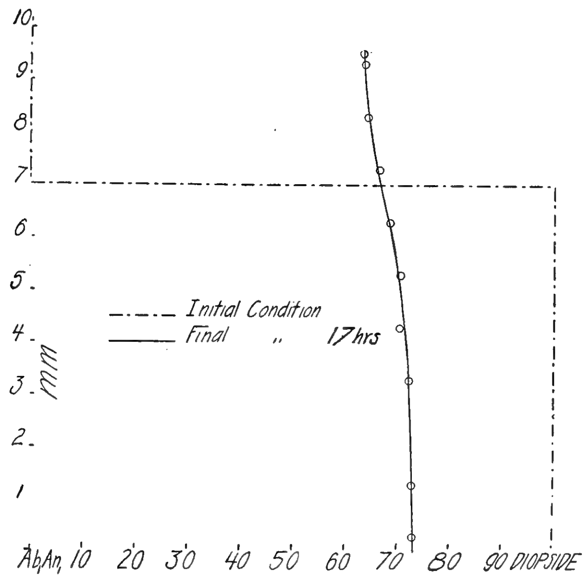

FIG. 4.—Diffusion experiment No. 37

agreement with theory if the variation of the coefficient of diffusivity is taken into account. It is possible, too, that from the observed values a definite quantitative relation between the coefficient of diffusivity and the concentration could be calculated, but the writer has not been able to find any attack upon a problem of this

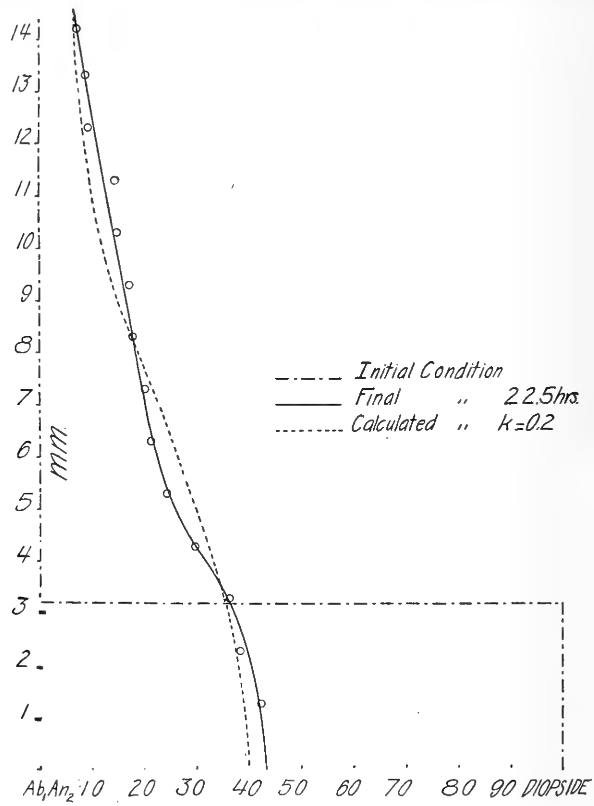

FIG. 5.—Diffusion experiment No. 25

kind in the various treatments of diffusion of concentration or of temperature. All of these adhere to a constant value of the coefficient of diffusivity. While in some respects it is highly desirable to check the present results more thoroughly along the lines indicated above, yet, for the purposes of the present paper, this is unnecessary. The results afford us very definite information as to the

magnitude of diffusion in silicate melts, which was the objective in mind when the work was undertaken.

VALUES OF DIFFUSIVITY

We cannot speak of a diffusivity "constant" in connection with the present results, but we may take the amount of diopside which penetrates to the surface layer as an indication of the average diffusivity of diopside in a liquid mixture of diopside and plagioclase. In this sense we find the "average diffusivity" of diopside in Ab_2An_1 mixture from Experiment No. 24 (Fig. 1), $k=0.015$ in cm.^{2} per day, in Ab_1An_1 mixture from Experiment No. 21 (Fig. 3), $k=0.14$, and in Ab_1An_2 mixture from Experiment No. 25 (Fig. 5), $k=0.2$. We therefore observe a progressive increase in the value of k with increase in the amount of anorthite in the diffusion mixture. Since it is well known that plagioclases rich in anorthite afford less viscous liquids than those rich in albite, we have further evidence of the increase of rate of diffusion with decrease of viscosity. If we compare the results of Experiment No. 21 (Fig. 3) and 37 (Fig. 4), we find again an increase in the value of the diffusivity, being $k=0.14$, and $k=0.3$, respectively.[1] In both these examples, however, the plagioclase Ab_1An_1 was used, the difference being that an increasing proportion of diopside was added. As a consequence, a higher value of the diffusivity is found for the mixture which was richer in diopside and therefore of lower viscosity. All of the results are therefore consistent with the assumption that the diffusivity varies inversely with the viscosity, which is in turn dependent on composition.

By way of comparison of the diffusivities here found with measured values of diffusivities for other substances, it may be noted that for common salt diffusing in water $k=1$ at 15°C., for gold in molten lead at 492°C., $k=3$, for solid gold in solid lead at 150°C., $k \doteq 0.0043$ in the same units as those used above. The

[1] In Figure 4 no calculated curve is shown for the reason that the calculated curve for $k=0.3$ sensibly coincides with the observed curve. Since the final result will be uniformity in all cases, whether the diffusivity varies with composition or not, a close approach of the calculated curve to the observed curve is to be expected in cases where diffusion is far advanced (Fig. 4). On the other hand, where the upper layers have not yet been affected, the greatest divergence between observed and calculated values is to be expected (Figs. 1 and 2).

diffusivities obtained for the silicates are, therefore, much smaller than those of salts in solution and those of molten metals, but much greater than those of solid metals. Some of the higher values obtained for the silicates are comparable with those of certain relatively viscous organic liquids. The diffusivity constant of glycerine in propyl alcohol at 17°C. is, for example, approximately 0.2.

APPLICATION OF RESULTS

It is not at all likely that the diffusivities of substances in mutual solution in rock magmas can be significantly greater than those determined for the plagioclase-diopside mixtures, and in many viscous magmas they would no doubt be considerably less. For the purpose of applying the results to diffusion problems in petrogenesis a value has been taken very close to the highest, viz., 0.25, which is at the same time convenient in calculations. The Soret action is one of the diffusion phenomena that has been considered of possible importance in magmas. It has been found in the laboratory that if a tube containing a solution is heated at one end and cooled at the other there is usually a concentration of the solute toward the cold end which depends upon the difference of temperature, the relative concentrations being, for many cases, inversely as the absolute temperatures. In cooling magmas the margin must be regarded as having a lower temperature than the interior, and there should presumably be a tendency toward a greater concentration of some substance or substances at the cooler margin. This introduces the possibility of composition differences in different parts of an entirely liquid magma, the differences being brought about by diffusion. If cooled entirely by conduction, the temperature of a magma brought into contact with cold country rock should at the border quickly assume a value midway between that of the magma and that of the country rock. For a long period thereafter cooling at the margin is very slow (see Fig. 6).[1] We may imagine that the original temperatures of the magma and that of the surrounding rocks are such that during this long period of maintenance midway between them the magma is still above its temperature of beginning of crystallization. Here we would have

[1] See also Lane, *Ann. Rept. Geol. Surv. Michigan for 1909*, Fig. 18, p. 152.

the most favorable case conceivable for the establishment of a composition gradient as a result of a temperature gradient according to the Soret principle. These conditions would evidently obtain either when the temperature of the magma was very much above the crystallization temperature, or when that of the surrounding rock was not very much below it, the latter being the more likely case. The magma intruded into hot surroundings, perhaps into a cognate intrusive not yet cooled, is, therefore, the most favorable

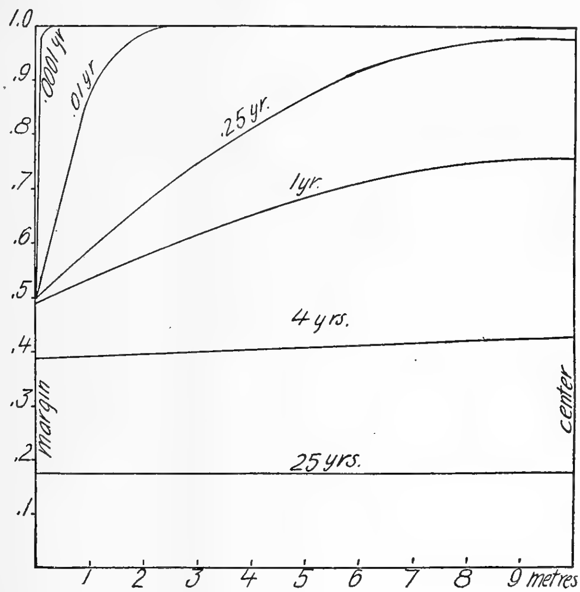

FIG. 6.—Curves of cooling of an intrusive igneous sheet 20 m. thick

subject for the working of the Soret action. Yet when we realize that the diffusivity of mass is, according to our determinations, from 10,000 to 100,000 times smaller than the diffusivity of temperature in rocks, it is apparent that the temperature of any igneous body will fall too rapidly to allow sufficient time for the Soret phenomenon to manifest itself.

This statement may perhaps be more readily appreciated if the Soret action is stated more definitely as a diffusion problem. In order to do so we may assume that the osmotic pressure is

proportional to the absolute temperature and that for this reason diffusion takes place until the concentration is inversely proportional to the absolute temperature. In other words, the effective concentration is, initially, inversely proportional to the absolute temperature, and diffusion takes place until the effective concentration is uniform. In applying these considerations to a cooling mass of rock, we may take for simplicity a tabular body. A solution of the problem of the cooling of such a body is given by the equation

$$\theta = \frac{\theta_0}{\sqrt{\pi}} \int_{\frac{-l-x}{2\sqrt{kt}}}^{\frac{l-x}{2\sqrt{kt}}} e^{-\beta^2} d\beta$$

where θ is the temperature at any point distant x from the margin, θ_0 the original temperature of the magma, the temperature of the wall rock being taken as zero and l is thickness of the intrusive.

If we take a tabular body of thickness 20 m. (i.e., 10 m. from center to margin) we may calculate the temperature in any plane at given distance from the margin at the end of any period of time. The results of such calculations are shown graphically in Figure 6, the curves representing the distribution of temperature at the end of various periods of time if the diffusivity is taken as 0.0118 in cm.^{2} per second. In this figure the temperature scale has no necessary absolute significance, 0 of the scale being merely the initial temperature of the surrounding rock and 1 of the scale being the initial temperature of the magma. It will be noted that at the end of one year the temperature at the margin is about halfway between the initial temperature of the surrounding rocks and that of the magma, while the temperature at the center is much higher. If it is supposed that the whole mass is still above its crystallization temperature, then the Soret action should be operative, that is, the effective concentration at any point should be proportional to the absolute temperature, and diffusion should take place until the effective concentration was uniform. Partly for simplicity and partly for the sake of obtaining an especially marked Soret effect we shall assume that the temperature scale of Figure 6 represents

absolute temperature, i.e., 0 of the scale is 0° absolute and 1 of the scale is 1000° absolute. After one year the temperatures at the margin and at the center are 490° and 760° respectively. Then the effective concentration at the center should be $\frac{76}{49}$ times that at the margin and diffusion should take place until the effective concentration is uniform, or until the real concentration at the center is $\frac{49}{76}$ times that at the margin. The problem is to find how long a time it would require for this diffusion to take place. Infinite time would, of course, be required to allow the process to go to completion, but we may find the time needed to give any assigned approach to this condition.

As a first step we may calculate the time necessary for the acquirement, from any arbitrary initial condition, of a concentration gradient represented by the curve showing the thermal gradient at the end of one year. This may be done by assuming a condition analogous to our experiments, viz., that all the material was first concentrated in a meter layer and by diffusion had acquired the gradient referred to. With the aid of equation (I) we find this to be very nearly true when $\sqrt{kt}$ in the limits of the integral has the value 500. If we take k as having a value close to the highest found in any of the experimental determinations, viz., 0.25, then $t=10^6$ days. But this is not the time we wish to know; it is that required to go on from this condition to practical uniformity. Again we find from the equation that practical uniformity (1:0.996) is obtained from the arbitrary initial condition when $\sqrt{kt}=1000$ or when $t_1=4\times10^6$ days. From this we get the desired time $t_1-t=3\times10^6$ days, or nearly 10,000 years. This shows that it would need about 10,000 years to obtain nearly the full theoretical Soret effect required by the curve of temperature distribution after one year in a mass of the dimensions chosen. In the meantime, as shown by the curves of Figure 6, the whole mass would have cooled to the temperature of the surrounding rocks. Even if we imagine it to be still above its crystallization temperature at the end of four years, it will be noted that most of the temperature gradient has been destroyed at that time so that

there is no reason for the action continuing even for four years. It should be observed that, although the times have been computed for a body of a definite size, the solution is really of a general nature, for if the body were n times as thick, the time on the one-year curve would be changed to n^2 years and the time required for the Soret phenomenon would be n^2 times as large. Nothing is gained, therefore, for the Soret effect by making the body larger or smaller.

In assuming that the scale of Figure 6 represents absolute temperature, we have, of course, taken an impossible condition for any body of rock. This would mean that the surrounding rocks were initially at 0° absolute and the magma had not yet begun to crystallize at 490° absolute = 217°C. The assumption was made on account of the convenience of referring both concentration and temperature gradients to the same curve, but even if we make reasonable assumptions as to the temperature of the magma and of wall rock, our conclusion will not be affected. We may even assume that the Soret effect for silicates is many times that deduced from the theoretical (absolute temperature) relation, yet the outstanding fact remains that the time required to produce a significant amount of concentration of material by diffusion is enormously greater than that required for the mass to cool off.

It may be noted that in speaking of a concentration toward the cool margin no mention has been made of what is concentrated. The reason is, of course, that it is not known. Ordinarily it is stated that the solute is concentrated toward the margin, but no distinction of solvent and solute can be applied to magmas. In conclusion, then, it may be stated that no concentration of any substance toward the cool margin could occur in appreciable amount in the time available for such action in a cooling mass of completely molten rock.

DIFFUSION TOWARD MARGIN DURING CRYSTALLIZATION

There is, however, another case of diffusion of material toward the cool boundary for which we know the nature of the material that should move in that direction. Harker lays stress on the fact that in any cooling mass of magma there should be a time

when crystallization of an early-formed mineral A takes place only near the border, the rest of the mass being still above the temperature of crystallization. The greater concentration of the substance A in the interior of the mass where none has yet precipitated should constitute a driving force tending to cause that material to diffuse toward the margin. This case may be stated fairly simply as a definite diffusion problem. When the temperature of a thin layer at the margin has fallen to such a value that a certain fraction of the amount of substance A in that thin layer has precipitated, then the magma in this layer is saturated with A. If this condition is maintained for an infinite period of time, the whole body of magma will eventually acquire the same concentration in A as this marginal saturated solution[1] and all of substance A in excess of this concentration will be precipitated at the margin. If we assume the contact surface plane as in a tabular mass, this is essentially the same problem as the heat-conduction problem involved in the cooling of a sheet of metal one face of which is kept at constant temperature. A solution of the problem is given by the equation:

$$c=\frac{2c_0}{\sqrt{\pi}}\int_0^{\frac{x}{2\sqrt{kt}}} e^{-\beta^2}d\beta$$

which gives the concentration c in terms of the original concentration c_0 at any point at distance x from the margin after the time t, the concentration of the marginal saturated solution being taken $=0$.

Values of c for various values of x and t have been calculated for a diffusivity 0.25 in cm.^{2} per day, approximately the highest experimental value, and are plotted as concentration curves in Figure 7. The figure shows that after two-thirds of a year the precipitating effect has been felt for a distance of 0.33 m. from the margin, all the rest of the magma being entirely unaffected. After sixty-four years the precipitating effect has been felt for a

[1] Neglecting the Soret effect.

distance of 3.3 m. while all the rest of the magma is unaffected, and similarly for the other concentration curves.

We can, moreover, determine from the figure the amount of material that would be precipitated at the margin at the end of any period of time. At the end of infinite time the concentration curve would correspond with the axis of abscissae, that is, the concentration at all points is zero of the scale, or equal to the concentration of marginal saturated solution. The whole area

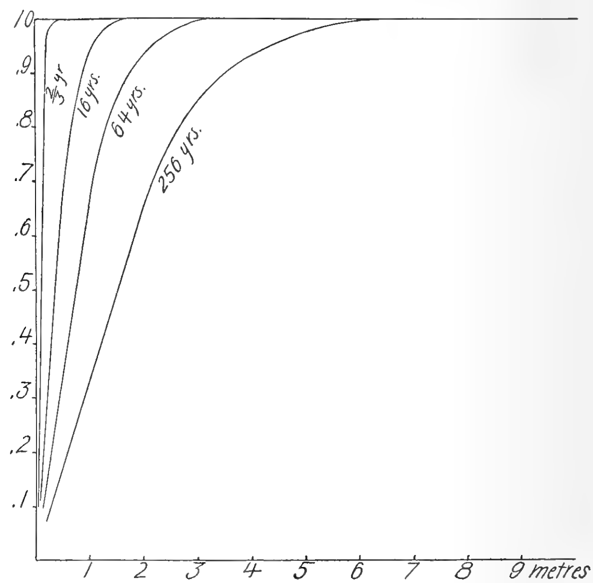

FIG. 7.—Curves of concentration in an igneous mass showing diffusion of material toward a cool boundary.

of the rectangle between the extreme ordinates of the figure represents, therefore, on a certain arbitrary scale the amount of material which would be removed in bringing the whole magma to the concentration of the marginal saturated solution. On the same scale the area lying above and to the left of the concentration curve for any time represents the amount of material removed during that time. We may take a specific case and imagine that the mineral precipitated at the margin is, say, amphibole, which occurs in the magma to the extent of 20 per cent, and that the

cooling of the marginal layer had proceeded until one-fourth of the amphibole contained in that layer (or 5 per cent of the layer) had been precipitated. If this condition were maintained the precipitation of amphibole at the margin would proceed by diffusion from the parts not yet cooled below the temperature of precipitation of amphibole. At the end of two-thirds of a year a proportion of the excess amphibole would be precipitated equal to the area between the one-year curve and the axis of ordinates. Regarding this area as a triangle of base sensibly 0.3 m. and assigning any arbitrary total thickness x to the mass, then the thickness of the amphibole deposit in meters would be

$$\frac{0.3}{2} \cdot \frac{1}{x} \cdot \frac{5}{100} x .$$

That is, a deposit 0.0075 m. or $\frac{3}{4}$ cm. thick would be formed on the margin in two-thirds of a year. Its thickness is independent of the total thickness of the mass of magma and all of it would be derived from a layer of magma 33 cm. thick. After sixteen years the deposit would be about 2 cm. thick, all coming from a layer of magma about 1.5 m. wide. After two hundred and fifty-six years the deposit would be about 8 cm. thick, all from a border portion less than 7 m. wide.

It is apparent that the possible results of diffusion after the manner postulated are exceedingly small. A mass of magma large enough to remain in the necessary condition for two hundred and fifty-six years would have a border phase 8 cm. thick. By necessary condition is meant that the margin should be cooled within its crystallization range and the main portion of the mass be not yet so cooled. The indications are that the mass would require to be at least 300 m. thick and be intruded under special conditions of temperature of magma and of wall rock, and the border phase would then be insignificant. Even by making more liberal assumptions as to the amount of chilling at the margin, say, a chilling sufficient to precipitate 25 per cent of the magma solution, the possible border phase would be increased in magnitude only five times. Moreover, as one increases the extent of marginal chilling, a stage is soon reached where so much precipitation takes

place in the marginal phase that no diffusion into that region can occur. One then arrives at a method of formation of a border phase that has been suggested by Daly, who regards the border phase as a chilled phase having the composition of the original magma.[1]

FORMATION OF REACTION RIMS

We have seen in the foregoing that the movement of large quantities of material through long distances by diffusion in a magma cannot be credited when the relatively rapid rate at which the magma must cool is considered. On the other hand, diffusion through short distances is to be expected, and such phenomena as the formation of reaction rims about foreign inclusions are readily to be attributed to diffusion. At the same time it should be noted that a rather wide reaction border will require a very considerable period of time for its formation if diffusion alone is active. Figure 7 enables one to form an idea of the period of time required for the diffusion of material from an inclusion to various distances in the surrounding medium if the scale of concentrations is reversed, that is, if zero is placed at the top and one at the bottom. The solution is by no means a rigid one for a small inclusion, but for a large slablike inclusion is sufficiently good to enable one to draw general conclusions. The figure shows that after sixty-four years the effect of the inclusion is barely felt for about 3 m. and is strongly felt (one-half saturation) for not more than 1 m. These considerations suggest that the formation of reaction rims up to 2 m. thick, such as those described by Ussing, about inclusions of quartzite in augite syenite at Kangerdluarsuk would require a period of time of the order of magnitude of one hundred years if diffusion alone were operative.[2]

The growth of crystals is itself largely dependent upon diffusion, but no quantitative estimate of the rate of growth is possible without some knowledge of the concentration gradient along which flow of material takes place, that is of the degree of supersaturation possible in the liquid interstitial to the crystals. The

[1] *Igneous Rocks and Their Origin*, p. 237.

[2] *Geology of the Country about Julianehaab, Greenland*, p. 362.

fact that rocks are normally millimeter-grained rather than centimeter- or meter-grained even in large masses is, however, a tribute to the slowness of diffusion in magmas. The fact that certain conclusions are reached above on the assumption that diffusion acts alone should not be taken as indicating that the writer believes that no other processes could occur. To account for the extremely coarse grain of many pegmatites, for example, it seems necessary to assume circulation of solutions, and in many other cases cited circulation (convection) would be inevitable.

SUMMARY

The rate of diffusion in certain silicate melts has been determined experimentally by permitting diffusion against gravity of a heavy liquid into a lighter liquid. The concentration curves found are not coincident with any theoretical curves calculated on the basis of a constant value of the diffusivity, but can be interpreted on the assumption that the diffusivity varies with concentration and is less for concentrations corresponding to more viscous liquids than for those corresponding to less viscous liquids. Taking as representative of the "average diffusivity" the amount of material which penetrates into the upper layer, the following values of the average diffusivity (k) were found: for diopside into Ab_2An_1, $k=0.015$; for diopside into Ab_1An_1, $k=0.14$ to 0.3, depending on the proportions; and for diopside into Ab_1An_2, $k=0.2$, all in cm.^{2} per day.

The value 0.25 (close to the maximum experimental value) is taken as probably representing a fair estimate of diffusivity in magmas, and with this as a basis it is shown that such phenomena as the formation of border phases about large bodies of igneous rock by diffusion cannot be considered possible in the time available for such action in a cooling magma. On the other hand, the formation of reaction rims about inclusions may be attributed to diffusion, though for very wide rims a considerable period of time will be required.

THE PHYSICAL CHEMISTRY OF THE CRYSTALLIZATION AND MAGMATIC DIFFERENTIATION OF IGNEOUS ROCKS

J. H. L. VOGT
TRONDHJEM, NORWAY

INTRODUCTION

In a later paper of this treatise (I–IV) are given the physico-chemical laws which govern the *crystallization* of igneous rocks. Subsequently, it will be shown that the same laws can be applied to the explanation of the *chemical composition* of igneous rocks, and consequently also of *magmatic differentiation.*

As I shall often refer to my earlier publications on the problems here discussed, I give a list of those of most importance:

"Studier over slagger," *Svenska Vet.-Akad. Handl.*, 1884. (Stockholm, 1885.)

"Beiträge zur Kenntnis der Gesetze der Mineralbildung in Schmelzmassen und Ergussgesteinen," *Archiv for Mathem. og Naturv.*, Vols. 13 and 14. (Kristiania, 1888–90.)

"Die Silikatschmelzlösungen," I and II. *Kristiania Videnskabs-Selskap*, 1903, 1904.

"Physikalisch-chemische Gesetze der Krystallisationsfolge in Eruptivgesteinen," *Tschermaks min. und petrogr. Mitt.*, Vols. XXIV (1905), XXV (1906), and XXVII (1908).

"Über anchi-monomineralische und anchi-eutektische Eruptivgesteine," *Kristiania Vid.-Selsk.*, 1908.

"On Labradorite-Norite with Porphyritic Labradorite-Crystals: a Contribution to the Study of the "Gabbroidal Eutectic," *Quart. Jour. Geol. Soc.*, 1909.

"Über das Spinell: Magnetit-Eutektikum," *Kristiania Vid.-Selsk.*, 1910.

"Die Sulfid: Silikatschmelzlösungen" (a review, 97 pages), *Norsk Geologisk Tidsskrift* (Kristiania), IV (1917).

"Die Sulfid: Silikatschmelzlösungen. Part I. Die Sulfidschmelzen und die Sulfid: Silikatschmelzen." *Kristiania Vid.-Selsk.*, 1919. Later will appear Part II. Die Nickel-Magnetkies-Lagerstätten.

For geological surveying, chemical analysis, photographs, etc., for this publication I have had contribution from Den Tekniske Hoiskoles Fond (The Foundation of the Technical University of Norway).

ERRATUM

In the article by J. H. L. Vogt, p. 318, in the first line, delete the words "a later paper of."

I

REVIEW OF THE PHYSICAL CHEMISTRY OF THE CRYSTALLIZATION OF IGNEOUS MAGMAS

In the examination of these laws two different methods can be used: (*a*) the *synthetic*, in which there is an opportunity for precision-determinations, especially of temperature. Previous investigations on the crystallization of silicates from melts have been, nearly without exception, conducted at the pressure of one atmosphere; (*b*) the *analytic*, mainly based on the study of the structure of the rocks. In this manner we may examine the sequence of crystallization, and so also the "individualization-fields" of the minerals, further the mix-crystal systems, the chemical composition of eutectic intergrowths, etc.—all under the physical conditions, especially with regard to pressure and time, present during the solidification of the different igneous rocks.

The synthetic method forms the important base. The analytic method gives us, in particular, information as to the extent to which the results of investigations conducted chiefly at atmospheric pressure and during short periods of time, can be transferred to apply to the physical conditions under which the crystallization of magmas took place.

The two methods, therefore, go hand in hand and complete each other.

REMARKS ON THE STRUCTURAL CRITERIA FOR THE SEQUENCE OF CRYSTALLIZATION

The sequence of crystallization in igneous rocks may usually be determined by the complete, partial, or wanting idiomorphism of the minerals, by the inclusions, by deposition on a solid body (*Fixkörper-Absatz*), by "together-swimming" structure (*synneusis-struktur*, see below), by law-governed intergrowths, etc.

The complete *idiomorphism* of a primary mineral A against all the other minerals shows that its crystallization was finished before the commencement of the solidification of the others. From the partial idiomorphism of the primary mineral A against the primary mineral B whose idiomorphism is wanting, we can infer that the crystallization of A had commenced at an earlier stage than the crystallization of B. But we must not draw the more extensive

conclusion that A in its entirety had crystallized before the commencement of the crystallization of B. As to the conceptions allotriomorphism, hypidiomorphism, and panidiomorphism, we refer to the petrographic textbooks.

Inclusions of an idiomorphic, primary crystal A in B implies that A had crystallized earlier than the surrounding parts of B. But if A only appears in the exterior zone of B, the interior part of B may have crystallized earlier than A. And even if inclusions of idiomorphic crystals of A appear evenly distributed over the whole of B, also in the kernel of B, it may be that part of A also may have crystallized at a later stage. As an example, idiomorphic crystals of apatite, as is known, in many cases appear in the oldest silicates and in the ore minerals, indicating that the apatite crystallized before the commencement of the solidification of the iron ore and the silicates. But I warn against the conclusion, which is often drawn, that the apatite in its entirety crystallized during the earlier stage.

Further, it must be taken into consideration that small portions of the mother-liquid occasionally may be inclosed or included in a mineral during its growth. As example we refer to the well-known zonally arranged glass-inclusions in leucite, sanidine, etc., in many dyke and effusive rocks. If corresponding magma-inclusions occur in deep-seated rocks, a complete crystallization of this material will take place. Thus the result will be inclusions in the host of a later-crystallized mineral.

Inclusions of mineral A in mineral B may furthermore be due to the fact that A originally, at high temperature, occurred as a solid solution in B, and that afterward, owing to reduced solubility by decreasing temperature, A separated from the solid solution. As an example we refer to the well-known inclusions of perthitic albite or albite-oligoclase in the microcline of granites, etc. The microcline (or orthoclase) dissolved about 28 per cent Ab+An, the greater part of which later separated during refrigeration. Further may be mentioned the secretion of lamellae of monoclinic pyroxene in orthorhombic pyroxene[1] and conversely also of ortho-

[1] Cf. the general account in my publication in *Tscherm. min. u. petrogr. Mitt.*, Vol. XXIV (1905), pp. 537–42.

rhombic pyroxene in monoclinic.[1] In the same manner the well-known microscopic inclusions, often with idiomorphic contour, of titanic iron ore in hypersthene, diallage, and plagioclase of gabbros may be explained.[2] The latter inclusions were often interpreted by earlier investigators as older than the host-mineral, but in their present form they must be explained as later secretions from an originally solid solution. As may be understood from this account concerning the inclusions we must take into critical consideration a great number of momentums in determining the successive age of the minerals.

When a substance is segregated from a solution, it often, as is well known, adheres to a solid already present (solid body or *Fixkörper*). The result of this is the *deposition on a solid body* (*Fixkörper-Absatz*), which is also very important in the solidification of igneous rocks. We may here, for instance, refer to Figure 34, illustrating the deposition of spinel on pyrite; to Figure 33, illustrating the deposition of titanomagnetite on olivine; and to Figure 35, where in one place pyrite has been deposited on apatite while in another apatite has been deposited on pyrite.

The individuals of a mineral, segregated from a magma at an early stage, frequently swam together to assemblings or aggregates, the result of which is a structure, for which I propose the term *together-swimming structure* or *synneusis structure*.[3]

This together-swimming may occur very rapidly. I refer to my publication "Die Sulfid: Silikatschmelzlösungen" (I, 1919), Figure 11, illustrating assemblages of octahedrons of magnetite in a bessemer-matte, consisting chiefly of Cu_2S, and to Figure 28, *a*, illustrating assemblages of small individuals of zincblende in a slag. The solidification period of the two molten masses just mentioned, respectively molten sulfide and molten slag, needed only a very short time, at most half an hour.

[1] W. Wahl, "Die Enstatitaugite," *Tscherm. min. u. petrogr. Mitt.*, Vol. XXVI (1906).

[2] In this connection we refer to a treatise by A. Johnsen, "Regelmässige Einlagerung von Eisenglanz in Cancrinit," *Centralbl. f. Min., Geol. und Pal.*, 1911, and by O. Andersen, "On Aventurine Feldspar," *Amer. Jour. Sci.*, Vol. XL (1915).

[3] Composed of *σύν*, *syn* = together and *νεῦσις*, *neusis* = swimming.

The phenomenon here discussed may, as to igneous rocks, be illustrated by Figure 1, representing a dunite from the Hestmandö-field in the northern part of Norway, with an average of only about 1 per cent chromite. In some parts of the thin section chromite is entirely or almost entirely lacking, but in other places we may find aggregates of ten to twenty small octahedrons of

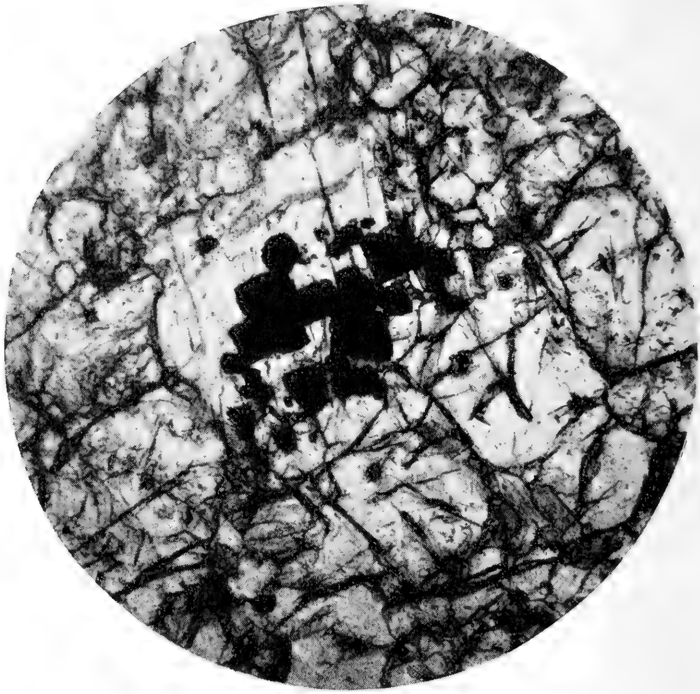

Fig. 1.—Dunite from the Hestmandö-field, northern Norway. Groups of octahedrons of chromite, illustrating "together-swimming" or synneusis structure. (Photo. 25:1.) (Black = chromite, white = olivine.)

chromite which, as well with reference to the idiomorphism against the olivine as with reference to the together-swimming structure, must have crystallized while the olivine was still in a molten condition. From this and other dunite rocks with a little chromite we may draw the conclusion that the chromite commenced crystallizing earlier than the olivine, when there was at least 1 or 0.5 per cent, perhaps only 0.33 per cent, chromite present. But

this does not exclude that the olivine may have commenced crystallizing earlier than the chromite when the latter only amounted, for example, to 0.1 or 0.05 per cent. The earlier silicate minerals, for example, olivine in olivine-rich gabbros, and hypersthene or diallage in hypersthene- or diallage-rich gabbros, also often show the together-swimming structure, as in Figures 10, 12, 13, 20 and 21, and 33.

The *relative commencement* of the solidification, especially of the minerals that commence crystallizing at a somewhat early stage, may often quite easily be decided by the structure. On the other hand, the allotriomorphism of a mineral C, against the minerals A and B, shows that C only commenced crystallizing after an often quite essential part of A and B had already solidified. Especially where the later mineral C is present in a small quantity, its allotriomorphism in connection with its appearance as *Zwischenklemmungsmasse* (or mesostasis) presents an easily recognizable criterion that it belongs to a very late stage of the crystallization. But at this late stage the minerals A and B will in many cases have continued forming. We refer to the explanation given in connection with Figures 17 and 18.

The *simultaneous crystallization* of two or more minerals may be manifested in various ways. With two simultaneously crystallizing minerals, each may grow until the individuals of A happen to collide with the individuals of B. Or some of the segregating mineral A may be deposed on the already solidified crystals of B, and some of the simultaneously segregating mineral B on the already solidified crystals of A.

In this manner we may observe crystals of plagioclase with quite good idiomorphic contour against the hypersthene or diallage and, further, crystals of hypersthene or diallage with quite good idiomorphic contour against plagioclase in the same thin section of an anchi-eutectic norite or gabbro. In the deep-seated rocks it may in such cases often be quite impossible to decide which of the two minerals first commenced crystallizing.

It may also often occur that the two minerals crystallize in an intimate intergrowth. In this manner simultaneously crystallizing

feldspar and quartz, as is well known, sometimes produces micropegmatitic or granophyric structure, as in graphic granite. Corresponding structure, which gives an evidence of a crystallization along a eutectic boundary curve, or exceptionally by a binary (or ternary or still more complex) eutectic, is also sometimes found with other minerals, for instance, between olivine and magnetite (see Fig. 28).

In many, possibly in most, cases the crystallization of the minerals A, B, C, etc., takes place in the following manner: Each mineral begins crystallizing at its proper stage, and continues to grow until the entire magma has solidified. As an extreme example we may choose apatite. This phosphate is only slightly soluble in silicate magmas at temperatures just above that at which the silicates commence to crystallize. If there is 0.20 per cent apatite present, the essential portion, perhaps 0.18, 0.19, or 0.195 per cent, has already crystallized before the silicate minerals have commenced segregating. But we may be pretty certain that a trifle phosphate, 0.02, 0.01, or perhaps only 0.005 per cent, still exists in solution at this stage and little by little solidifies later. It has, however, not been possible for me to substantiate this by observation with respect to the apatite; but I have been able to establish that spinel, when present only as 0.01 or at most 0.02 per cent, only commenced crystallizing after a great part of the silicate mineral A had solidified. (See Fig. 33 and the chapter on spinel.)

Fe_3O_4 and the different ferromagnesian silicates are only slightly soluble in acid—or granitic—magmas, and therefore commenced crystallizing at an early stage. We find, however, as is discussed below, a small remnant of magnetite and ferromagnesian silicate in the final product of the solidification. We may consequently draw the conclusion that the essential part of the magnetite and the ferromagnesian silicate was certainly solidified during the first stage of the crystallization, but that a little remnant stayed in the solution and was solidified later.

In a binary system, type IV, of two discontinuous mix-crystals—A, melting at relatively high temperature (for example, FeS_2,

at high pressure), and B, melting at relatively low temperature (for example, FeS)—idiomorphic crystals of B never appear in A, but, on the contrary, idiomorphic and usually somewhat resorbed crystals of A appear in B. Here the sequence of crystallization, without regard to the proportion of weight between A and B, is first A and later B, and this to be understood thus, that the crystallization of A was *completed* before that of B began. (See Fig. 36.)

In the usual silicate eruptives, most frequently consisting of a whole series of components, we may also meet corresponding crystallization of a certain mineral, completely solidified at a relatively early stage. See Figures 20 and 21, representing crystals of hypersthene imbedded in diallage.

Another case of crystallization *completed* at an early stage is illustrated by Figures 8 and 9 (and a theoretical explanation given below) with crystallization at the beginning stage of hypersthene, while the Fe-Mg silicate at a later stage entered into biotite.

In this treatise (Part I) we are only going to consider the solidification of the rocks (the transition from liquid to solid phase). We, however, also discuss the continued change of the minerals, which may be founded on the later crystallization of a substance originally in solid solution, and furthermore we are going to deal with the reactions which appear in the solid phase on the boundary between two minerals, and which are an immediate result of the cooling of the rocks after completed crystallization. We shall, however, not discuss the later changes, which are not a direct result of the solidification of the rocks, but are founded on exterior incidents, as, for instance, dynamo and contact metamorphosis, chemical actions, etc.

INTRODUCTORY REMARKS CONCERNING THE APPLICATION OF THE PHYSICO-CHEMICAL LAWS TO THE CRYSTALLIZATION OF MAGMAS

Magmas usually consist of a whole series of components, which entail a complication of the equilibrium existing in the magma and consequently also of the laws of the crystallization.

In many cases, however, an essential simplification of these complications takes place, as, according to H. E. Boeke,[1] the

[1] *Grundlagen der physikalisch-chemischen Petrographie* (1915), p. 104.

following assertion is applicable: "The saturation-boundary between binary mix-crystals as well as in general the equilibrium between two solid phases of a binary system do not change by contact with other phases and components, when these new entering components do not form solid solutions or stoechiometric compounds with the former solid phases." The proportion between An and Ab in the segregated plagioclase mix-crystal will in this manner be the same whether the crystallization takes place in a pure An+Ab melt or in a silicate melt (or magma) which besides plagioclase also delivers, for instance, magnetite, olivine, etc.

In order to investigate the laws of crystallization of the principal components of the magma, we may generally leave the components which are present only in subordinate quantity out of consideration, provided that the latter do not form solid solutions or enter into mix-crystal combinations with the principal components. We must, however, take into consideration that when A, B, and C form a ternary eutectic, and C only is present in small quantity, the simultaneous crystallization between A and B, along the eutectic boundary between A and B, will not be identical quantitatively with the composition of the binary eutectic between A and B. If C, however, is present in minimal quantity, the difference between the point in question on the eutectic boundary between A and B and the binary eutectic A:B will be so inconsiderable that it practically may be left out of consideration.

THE FELDSPARS, AB:AN, OR:AB+AN.

The binary system *Ab:An* (with melting-point Ab = 1100±10° and An = 1550°±2°) belonging to mix-crystal type I has been studied in detail by N. L. Bowen[1] of the Geophysical Laboratory of the Carnegie Institution of Washington at the pressure of one atmosphere (and with chemically pure substances). We reprint Bowen's graphic exhibit as Figure 2, where the great horizontal difference between the liquidus and solidus curves is shown.

As an example we may mention that from a molten mass, Ab_1An_1, the first mix-crystal, separating at 1450° (without super-

[1] *Amer. Jour. Sc.*, Vol. XXXV (1913).

saturation), has a composition nearly exactly $Ab_{20}An_{80}$, or Ab_1An_4. Even before Bowen's investigation (1913), I had (see especially *Tscherm. Mitt.*, Vol. XXIV [1905]), on the basis of the well-known zonal structure of the plagioclases, and further on the basis of the proportion between Ab:An in the total plagioclase calculated from the analysis of the rocks and in the first segregated crystal, decided that the system Ab:An belongs to type I, and, furthermore, that in the crystallization of plagioclase in eruptive rocks, especially in

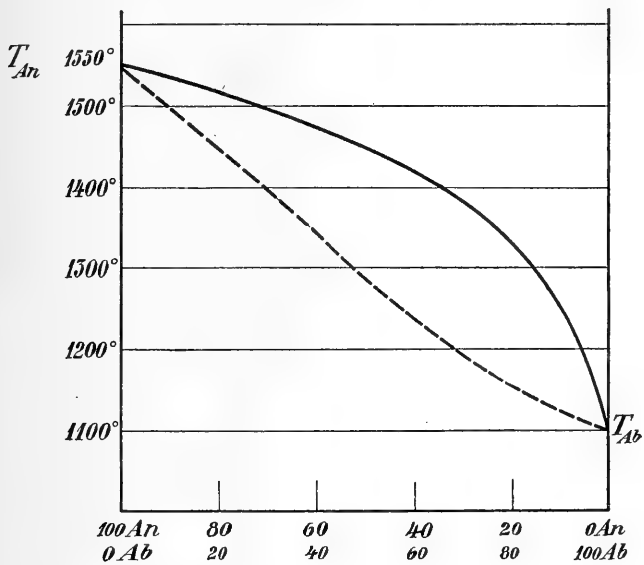

FIG. 2.—The melting-diagram, An:Ab, after Bowen

dike and effusive rocks, there is a very great horizontal difference between the liquidus and the solidus curves.

The binary diagram for Ab:An, at the pressure of one atmosphere and for 100 per cent Ab+An, may with unessential modifications of the horizontal difference between the two curves (see a following chapter) be transferred to magmas, crystallizing at high pressure, which besides Ab and An contain other components. The investigation of the mix-crystal system and the proportion between the liquidus and solidus curves for Ab:An may consequently also be applied to rocks, which besides plagioclase also

contain a great quantity, 40 per cent and still more, of foreign components, as quartz and Fe-Mg silicates. Petrographic experience proves that in such rocks we can detect no difference in the horizontal distance of the two curves, and this is in best accordance with the general law cited on page 326.

Or:Ab or Or:Ab+An.—Because of the extreme viscosity which characterizes melted $KAlSi_3O_8$ and $NaAlSi_3O_8$ (without or with only a small quantity of $CaAl_2Si_2O_8$), the synthetic study of the system Or:Ab or Ab+An is connected with exceptionally great difficulties. We may therefore here use the analytic method, cf. my earlier publication in *Tscherm. Mitt.*, Vol. XXIV (1905).

When Or is predominant, orthoclase first crystallizes, and when Ab+An is predominant, plagioclase first. The boundary ("individualization-boundary") is decided by the following method of investigation:

In a number of rocks, where the proportion Or:Ab:An in the entire rock was determined on the basis of the rock analysis, we find crystallized as No. I: *plagioclase* at 32, 32, 32.5, 32.5, 33.5, 34, 36, 36, 37.5, 39.5, 40, and 41 Or:Rest Ab_{+An}; *orthoclase* (or microcline) at 42, 43, 43.5, 43.5, 46, 47, 47, 50, 50, 50.5, 52, and 52.5 Or:Rest Ab_{+An}.[1] Ab_{+An} here represents Ab with a small but variable quantity of An, consequently albite, oligoclase, and andesine.

Between the two "individualization-fields" lies the boundary for orthoclase: albite, oligoclase, or andesine at about 0.4 Or:0.6 Ab or Ab_{+An}, perhaps nearest 0.42 Or:0.58 Ab_{+An}, applying to magmas at high pressure and with predominant Or+Ab+An with some mixture of other components. This boundary must be interpreted as an eutectic boundary curve.

When orthoclase (or microcline) crystallizes at *high* temperatures from granitic magmas, which besides predominant Or also contain a good deal of Ab_{+An}, there enters in the orthoclase up to about 30 per cent (or 28 per cent) of Ab_{+An}, which is partially separated by the cooling of the solid solution as albite or albite-oligoclase-

[1] Two uncertain determinations with 38 and 36 Or are here left out of consideration.

perthite. By the crystallization of acid plagioclase, the latter absorbs up to about 10 per cent (or 12 per cent) Or, which on cooling may give anthiperthite. Basic plagioclase (labradorite-bytownite) seems to absorb a somewhat smaller quantity of Or. Concerning this matter we refer to the chapter on the anorthosites in Part II.

By the above-mentioned crystallization in the acid magmas of Or and Ab (or Ab_{+An}) there result two minerals: orthoclase about 0.7 Or + 0.3 Ab (or Ab_{+An}), and acid plagioclase, about 0.9 Ab (or Ab_{+An}) + 0.1 Or.

Orthoclase and albite have almost exactly the same melting-point, and almost exactly the same atomic weight, probably also about the same latent melting-heat, etc. Granted a binary eutectic system (type V), the eutectic must lie almost exactly midway between 0.7 Or + 0.3 Ab and 0.9 Ab + 0.1 Or, consequently at 0.4 Or:0.6 Ab, just as we found above.

As An has a far higher melting-point than Or (and also than Ab), it is likely that the eutectic boundary between orthoclase and plagioclase with decreasing Or is characterized by increasing An in the plagioclase. In this manner we shall probably find the individualization-boundary between orthoclase and labradorite at 0.45—0.5 Or:0.55—0.5 Ab_{+An}.

This explanation is, however, not explicit, as we have not taken into consideration that in certain rocks or under certain conditions, the details of which we are not acquainted with, the two independent minerals, orthoclase and albite (or some other acid plagioclase), do not appear, but instead we find the mineral anorthoclase.

In my treatise in *Tscherm. Mitt.*, Vol. XXIV (1905) I pointed out the fact that a whole series of analyses of anorthoclase shows relations about 0.4 Or:0.6 Ab (or about 0.42 Or:0.58 Ab_{+An}) and I set forth the hypothesis that the anorthoclase might be defined as a microscopical or submicroscopical eutectic intergrowth of orthoclase (microcline) and albite (or some other acid plagioclase). This hypothesis is, however, quite dubious, and the physicochemical interpretation of the anorthoclase is still an open question.

QUARTZ (QU) AND FELDSPARS. QU:OR, QU:AB, QU:AN, QU:AB+AN, QU:OR:AB+AN

The binary system SiO_2:$CaAl_2Si_2O_8$ (*An*) has been examined by G. A. Rankin and Olaf Andersen[1] (at the Geophysical Laboratory, Washington) with the result: An, melting-point = 1550° ± 2°; Eutectic SiO_2:An = 52 per cent An:48 per cent SiO_2, melting-point 1353° ± 2°.

According to I. B. Ferguson and H. E. Merwin[2] (1918) the melting-point for tridymite is 1670 ± 10°C; for cristobalite 1710 ± 10°C. K. Endell and R. Rieke[3] (1912) decided for cristobalite 1685 ± 10°. N. L. Bowen[4] decided a somewhat lower temperature, and C. N. Fenner[5] thought he might fix the melting-point of cristobalite at 1625°, which, however, according to the latest precision-investigations, must be a little too low.

It is a matter of course that *Qu and Or*, as well as *Qu and Ab*, in the same manner as Qu and An, must form a binary eutectic. Because of their extreme viscosity the binary eutectics Qu:Or and Qu:Ab are not experimentally determined. We may therefore here use the analytic method.

We shall commence with *graphic granite*, the structure of which, as already established many years ago by W. C. Brögger,[6] is due to a simultaneous crystallization of quartz and feldspar. That is to say, the crystallization took place at a eutectic point or along a eutectic boundary curve.

Referring to my earlier publications[7] on the problem in question, we are going to give a collocation of all the hitherto published usable or at any rate somewhat usable analyses of graphic granite from pegmatitic granite dikes.

[1] The System Anorthite-Forsterite-Silica," *Amer. Jour. Sci.*, Vol. XXXIX (1915).

[2] *Amer. Jour. Sci.*, Vol. XLVI (1918).

[3] *Zeitschr. f. anorg. Chemie*, Vol. LXXIX (1913).

[4] *Amer. Jour. Sci.*, Vol. XXXVIII (1914).

[5] *Ibid.*, Vol. XXXVI (1913).

[6] *Geol. Fören. Förh.*, Vol. V (1881), and *Zeitschr. f. Kryst. Min.*, Vol. XVI (1890), I, pp. 148–59.

[7] "Silikatschmelzlös.," II (1904), and *Tscherm. Mitt.*, Vol. XXV (1906).

TABLE I

	No.	SiO_2	Al_2O_3	Fe_2O_3	FeO	MgO	CaO	Na_2O	K_2O	H_2O	Total
				Microcline Graphic Granite							
Norway	1	74.04	14.40	Nil	Nil	Nil	0.33	2.01	9.36	Nil	**100.18**
	2	74.00	14.31	Nil	Nil	Nil	0.39	2.42	9.02	Nil	**100.14**
	3	73.82	14.44	Nil	Nil	Nil	0.35	2.45	8.90	Nil	**99.96**
	4	74.47	15.13	Nil	Nil	Nil	....	....	.....	....	......
	5	73.70	14.11	Nil	Nil	Nil	0.39	3.04	8.72	Nil	**99.96**
Sweden	6	72.41	14.51	0.30	....	Nil	0.13	2.19	10.09	0.28	**99.91**
	7	74.58	13.37	0.24	....	Nil	0.32	1.16	9.80	0.57	**100.04**
United States	8	73.89	13.75	0.26	....	Nil	Nil	2.10	9.00	0.24	**99.24**
	9	73.92	14.26	0.30	....	Nil	Nil	2.06	8.99	0.11	**99.64**
	10	72.76	15.44	(InAl_2O_3)	....	Nil	0.19	2.35	9.28	0.15	**100.20**
	11	72.80	15.07	0.26	0.21	Nil	Nil	3.35	7.92	0.30	**99.91**
				Oligoclase Graphic Granite							
Norway	12	76.8	14.2	Nil	Nil	Nil	1.7	6.1	1.5	....	**100.3**
Sweden	13	76.67	14.20	0.14	....	0.04	2.67	5.33	0.52	0.48	**100.04**

EXPLANATION

Nos. 1–5 and 12, see "Silikatschmelzlös.," II. No. 1 from Arendal; Nos. 2–3 from Hitterö; No. 4 from Raade; No. 5 from Arendal is industrially pulverized graphic-granite, in which a trifle surplus feldspar is not excluded. In all these analyses a small loss (0.1–0.3 per cent) caused by ignition is deducted. The precipitate of Al_2O_3 was in all cases entirely white, and on account of this a special analysis of iron was not made. Some iron, less than 0.1 per cent Fe_2O_3, is, however, not excluded. No. 12, approximate analysis from Evje. Nos. 1 and 5 were analyzed by A. Grönningsæter and E. A. Dalset, assistants at the time. Nos. 2–4 and 6 were analyzed by students. K_2O,Na_2O and CaO in No. 4 have not been included on account of less accuracy.

Nos. 6, 7, and 13 from A. Bygdén, *Bull. of the Geol. Inst. of Upsala*, Vol. VII (1906); from Elfkarleö, Skarpö, and Ytterby.

Nos. 8–10 from Edson S. Bastin, *U.S. Geol. Surv. Bull. 420* (1910); from Topsham, Me., and Redford, N.Y. (No. 10, a little Fe_2O_3 [n.d.] in Al_2O_3.) No. 11 from Hiriart Hill, Cal., analyzed by W. T. Schaller, cited by H. S. Washington, "Chemical Analysis of Igneous Rocks, 1884–1913," *U.S. Geol. Surv. Prof. Paper 99* (1907). (Cited below as Wash.)

In the above table I have not included: Analysis of oligoclase graphic granite from the West Indies in Bygdén's treatise (Wash., p. 267, No. 5, with 68.12 per cent SiO_2), as this specimen is greatly decomposed, with considerable new-formed epidote, etc. This analysis can therefore not be used in the determination of the proportion of quartz and feldspar.

Analysis No. 4 in E. S. Bastin's treatise (Wash., p. 108, No. 4, with only 71.00 per cent SiO_2), as Bastin informs us of this specimen: "Some small

areas of pure feldspar were associated with the graphic granite in this specimen, so that the silica percentage shown in the analysis is lower than it would be for graphic granite alone." The same is probably also the case of an analysis, by A. W. Howitt, 1888, from Victoria (Wash., p. 112, No. 30).

With respect to the other analyses which are noted in Washington's Index as graphic granite we remark: "The graphic microgranite," page 73, No. 2 (with 3.74 per cent Fe_2O_3, 2.81 per cent FeO), represents a rock, and not graphic granite. The same also applies to No. 19, page 94 (with 0.73 per cent Fe_2O_3, 0.78 per cent FeO, 0.99 per cent MgO).

In the treatise, above cited, by Bygdén, as also in a treatise by H. E. Johansson,[1] an analysis by P. J. Holmquist[2] has been taken as an example of an albite-graphic granite, showing 77.32 per cent SiO_2, 0.34 TiO_2, 11.62 Al_2O_3, 1.57 Fe_2O_3, 0.69 FeO, 0.10 MnO, 0.62 CaO, 0.80 MgO, 0.99 K_2O, 5.81 Na_2O, 0.65 H_2O, *total* 100.51. This was computed by Holmquist as: 39.0 per cent Qu, 49.3 Ab, 4.6 Or, 1.9 An, also 2.4 chlorite, 2.3 magnetite, 0.8 titanite, 0.3 water, and a little calcite. The analysis is from a thin dyke in diabase (Rödö) and does not permit any exact determination of quartz: albite in graphic granite.

The precision-determination of the quantitative proportion of quartz and feldspar is complicated partly because of the inevitable errors in the analyses, of which more below, and partly because we cannot always be certain that the analyses represent absolutely pure intergrowths of the two minerals. In granite-pegmatite dikes we sometimes meet specimens of which one part consists of pure feldspar, free from quartz, and the other part of graphic granite, retaining the crystallographic orientation of the feldspar. That is to say, some feldspar first crystallized alone, and later, having reached an eutectic boundary curve, it continued its growth simultaneously with quartz. Sometimes we may find in the center of a large specimen of graphic granite small parts of pure feldspar. In such cases, the analyses of course cannot be used for precision-determinations of the relative proportions of quartz and feldspar. We may here refer to Bastin's remarks concerning his analysis No. 4.

On the basis of the analysis, we are going to calculate the quantitative proportions of quartz and feldspar in our microcline-graphic granites, according to the following methods:

[1] *Geol. Fören. Förh.*, Vol. XXVII (1905).

[2] *Sveriges Geol. Unders.*, C. 181 (1899).

a) Originating from K_2O, Na_2O, and CaO we calculate the quantity $KAlSi_3O_8$ (Or), $NaAlSi_3O_8$ (Ab), and $CaAl_2Si_2O_8$ (An), and when this is done the sum of feldspar is deducted from the analytically determined sum of SiO_2, Al_2O_3, K_2O, Na_2O, and CaO. The difference is quartz. This method contains a very great source of error, as an error in the determination of K_2O, Na_2O, or CaO in the calculation of the amount of feldspar will be doubled respectively 6, 8.5, or 5 times. To this may be added the inaccuracy in the SiO_2 determination.

b) Originating from the proportion K_2O:Na_2O:CaO we calculate the percentage of SiO_2 in the feldspar (ex. 65.28 per cent in No. 1) and then we calculate the proportion between quartz (n) and feldspar ($1-n$), ex. for No. 1.:n. $100+(1-n).65.28=74.04$. A relative error in the determination of K_2O, Na_2O, or CaO will in this manner be eliminated. But an error in the determination of SiO_2 will be doubled nearly three times.

We have a control of the calculation, in method (*a*), in the calculated percentage of Al_2O_3 in the feldspar, compared with the percentage of Al_2O_3 found in the analysis. As an example, No. 5 shows 1.06 per cent too much Al_2O_3 in the calculation, consequently also too much feldspar, and so too little quartz. The calculated 21.58 per cent quartz must consequently be increased. On the other hand, No. 9 shows 1.12 per cent too little Al_2O_3 in the calculation, consequently too little feldspar or too much quartz.

With combined consideration of both methods of calculation, and of the sum of the analysis minus ignition and Fe_2O_3, we have under C written the probable percentage of the quartz. The rest is feldspar.

TABLE II

No.	A					B	C
	Qu	Or	Ab	An	Al_2O_3	Qu	Qu
1	25.60	55.51	17.05	1.64	−0.34	25.23	**25.4**
2	24.34	53.50	20.52	1.94	+0.16	24.90	**24.6**
3	24.54	52.78	20.78	1.74	−0.12	24.25	**24.4**
5	21.58	51.72	25.78	1.94	+1.06	23.52	**23.5(+?)**
6	20.55	59.84	18.57	0.65	+0.28	19.28	**20.3(+?)**
7	29.43	58.12	9.84	1.59	−0.25	27.53	**28.5(+?)**
8	27.01	53.38	17.82	....	−0.53	23.57	**25.3**
9	27.32	53.32	17.47	....	−1.12	23.67	**25.5**
10	22.94	55.04	19.94	0.95	−1.18	20.79	**21.9 (?)**
11	22.79	46.97	28.41	....	−0.97	19.19	**21.0 (?)**

The specimen No. 6 is an erratic block (with 0.30 per cent Fe_2O_3 and 0.28 per cent ignition). Here a minimal, hardly perceptible, decomposition is not excluded, so the determination 20.3 per cent quartz (or according to Bygdén 20.81 per cent) is probably too low. For No. 11 I have no supply of the original literature. The analysis shows 0.26 per cent Fe_2O_3, 0.21 FeO, and 0.30 ignition, and to the calculated result is added an interrogation point.

Arthur Holmes[1] has used another method in order to determine the quantitative proportions of quartz and feldspar in graphic granite from dikes of granite-pegmatite (in Mozambique), viz., Rosiwal's planimetric method. In this manner, from different localities, he found quartz amounting to 27.9, 27.1, 26.3, 25.6, 25.3, and 24.2, average 26.1 per cent (calculated in percentage by weight) and rest, 73.9 per cent of microcline with the ordinary perthitic admixture of albite-oblioclase. Consequently we have the following determinations of the quantity of quartz in microcline graphic granite:

Calculated from the quantitative analyses: 28.5 (too high ?), 25.5, 25.4, 25.3, 24.6, 23.5 (?), 21.9 (too low ?), 21.0 (?), and 20.3 (too low ?).

By the Rosiwal method: 27.9, 27.1, 26.3, 25.6, 25.3 and 24.2 per cent of quartz.

The majority of these determinations are subject to great sources of error, which may amount to several per cent. If we take this into consideration, I think I am justified in drawing the conclusions that the proportions of quartz and feldspar in microcline graphic granite from dikes of pegmatite are subject only to small variations or are practically constant, and that we may fix the proportion pretty closely at:

26 per cent quartz:74 per cent microcline.[2]

The graphic granite in granite-pegmatite dikes crystallized at a relatively late stage, viz., after the essential part of the mica and, most frequently, also a part of the feldspar had solidified. Only a trifle mica was left at the time for the forming of the graphic granite. But in addition to this, besides the components of the feldspars and the quartz, there was surely some H_2O present, possibly partly connected with SiO_2 in forming a separate component (as H_2SiO_3 [?]). The graphic granite will thus have crystallized from a solution, which consisted predominantly of the components of feldspar and quartz, but also of a little mica and a small quantity of a component, as H_2O and H_2SiO_3 (?). The graphic granite has in this manner crystallized at a eutectic boundary

[1] *Quart. Jour. Geol. Soc. London*, Vol. LXXIV (1919), p. 77.

[2] In "Silikatschmelzlös.," II (1904), I gave the proportion 25.75:74.25, which is practically the same.

curve, located *quite close* to the eutectic between the microcline components and quartz.

As microcline (from dikes of granite-pegmatite) always contains considerable Ab_{+An}, most often 25–30 Ab_{+An}:75–70 Or, it almost certainly has a somewhat lower melting-point than pure Or, and we must therefore assume that the eutectic Qu:Or contains a little more Qu than the eutectic Qu:microcline.

As the binary eutectic we shall assume 28 Qu:72 Or. As pure Ab has nearly the same melting-point as pure Or, nearly the same molecular weight (Or=279.4, Ab=263.3), and possibly also almost exactly the same melting-heat, it must be supposed that the eutectic Qu:Ab holds about the same percentage of Qu as the eutectic Qu:Or. As an approximation we may consequently assume 28 Qu:72 Ab as the binary eutectic.

QU:AN AND QU:AB, QU:OR

For the pressure of one atmosphere we have the synthetic determination: E_{Qu-An}=48 Qu[1]:52 An, at 1353°. Further we refer to the determinations just mentioned (for a very high pressure):

E_{Qu-Or}= *ca.* 28 Qu:72 Or
E_{Qu-Ab}= *ca.* 28 Cu:72 Ab.

Even if an error of a few per cent may be found in the latter statements, it is evident in any case that the binary eutectic Qu:Or or Qu:Ab contains much less quartz than the binary eutectic Qu:An. This is in accordance with Ab (as well as Or) having an essentially lower melting-point, in round numbers 450°, than An.

The course of the melting-curve (see Fig. 3) on the Qu side in the neighborhood of Qu will be about the same, whether the second compound is An or Ab (or Or). If we extend the curve, experimentally determined for E_{Qu-An} on the quartz side, and draw the curve on the feldspar side for E_{Qu-Ab} (respectively E_{Qu-Or}) about parallel with E_{Qu-An}, a binary eutectic E_{Qu-Ab} (respectively E_{Qu-Or}) will appear with composition about 25–30

[1] Qu here signifies christobalite from the melting-point to 1470° and tridymite from 1470° to 1353°.

Qu:75–70 Ab (or Or), just as we have in reality derived from the analyses of the graphic granites.

As the melting-points as well as the binary eutectics (as will be shown in a following chapter) are only very little displaced by pressure, we are justified in drawing a parallel between the eutectic SiO_2:An, determined for low pressure, and the eutectic SiO_2:Ab (or SiO_2:Or), calculated for high pressure. The case is somewhat

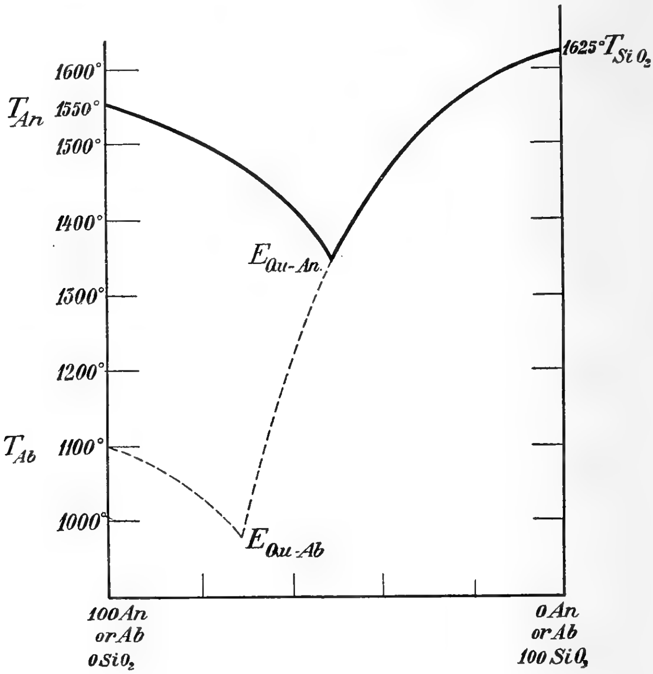

FIG. 3.—Melting-diagram, An:SiO_2 (after Bowen), and Ab:SiO_2 (schematic after Vogt).

complicated, however, by the fact that SiO_2 in one case is tridymite but in the other quartz.

The melting-point for the binary eutectic $E_{Qu\text{-}Ab}$ (or $E_{Qu\text{-}Or}$) must, according to the nature of the case, lie considerably lower than the melting-point of pure Ab (or Or), consequently considerably lower than 1100° and certainly somewhat lower than 1000°. As an estimate we set it at 975°, which should be approximately correct.

Qu:*Ab*+*An*.—In a *ternary* system consisting of two components (as, for instance, Ab and An), which form a binary mix-crystal system of type I, and a third component independent of the former (as Qu or $CaMgSi_2O_6$), there appear, according to F. A. H. Schreinemaker's theoretical investigations,[1] two melting-surfaces (Fig. 4),

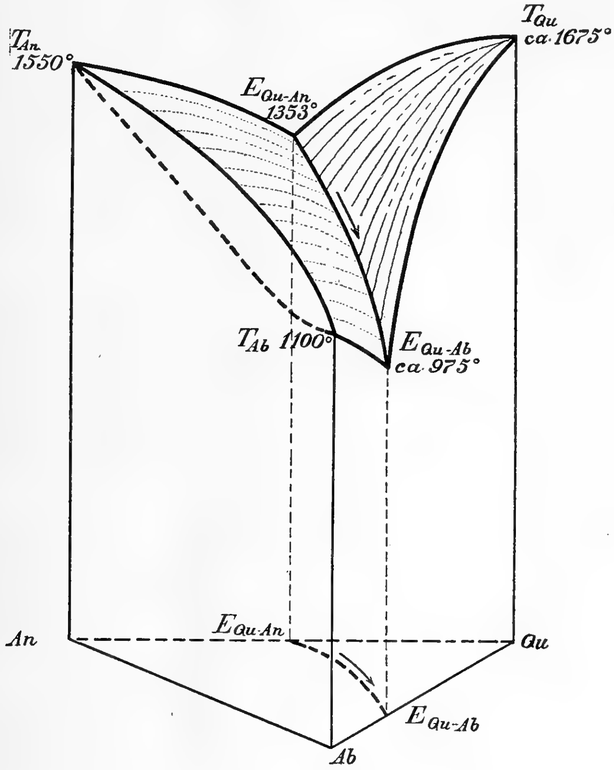

FIG. 4.—The ternary system An:Ab:Qu

which intersect in a curve, viz., "eutectic boundary-line"[2] or briefly a "eutectic line or curve." Three subcases may occur, accordingly as the eutectic boundary-line ($E_{Qu\text{-}An}$ to $E_{Qu\text{-}Ab}$ on Fig. 4) has a continuous decline, a minimum, or a maximum. In the ternary system Diops:Ab:An the eutectic line according to Bowen's experimental investigations (see Fig. 6) has a continuous decline from $E_{Diops\text{-}An}$ (at 1270°) to $E_{Diops\text{-}Ab}$ (at a little below

[1] *Zeitschr. f. physikalische Chemie*, 1905, Vols. 50, 51, and 52.

[2] This term I have used in my earlier treatise. Boeke (*loc. cit.*) uses the shorter term "eutectic line."

1100°), consequently with a difference of *ca.* 200° between the two points.

For the analogous system Qu:Ab:An, where the difference between the two points (E_{Qu-An} at 1350° and E_{Qu-Ab} at probably a little below 1000°) is still greater, certainly at least 350°, we may also suppose a continuous decline for the eutectic line. This line, on account of the steep decline near Ab of the binary liquidus curve between An and Ab, will probably assume the course outlined on the horizontal projection, Figure 5. The crystallization between Ab+An and Qu has consequently the same course as between Ab+An and diopside (see below). Even if the curve between E_{Qu-An} and E_{Qu-Ab}, contrary to our conjecture, should show a maximum in the vicinity of E_{Qu-An} or a minimum in the vicinity of E_{Qu-Ab}, this would in no degree worth mentioning modify the course of the curve in the horizontal projection.

We calculate the chemical composition of the end members and of a pair of intermediate compositions.

TABLE III

Percentage of	E_{Qu-Ab}	$E_{Qu-Ab+An}$ by		E_{Qu-An}
		$\frac{3}{4}$ Ab:$\frac{1}{4}$ An	$\frac{1}{2}$ An:$\frac{1}{2}$ Ab	
Qu.	28	36	42	48
Ab+An {Ab.	72	48	29	. . . .
Ab+An {An.	. . . .	16	29	52
SiO_2.	77.75	75.95	74.5	70.5
Al_2O_3.	13.95	15.2	16.25	19.0
CaO.	. . . .	3.2	5.85	10.5
Na_2O.	8.50	5.65	3.4	. . . .

That the calculation here given of the eutectic between quartz and albite, oligoclase, etc. (which is supported by the theoretical argument on the eutectic Qu:An at the pressure of one atmosphere, and by analogy conclusions according to the composition of microcline graphic granite) is essentially correct, is confirmed by the close conformity between the two analyses, Nos. 12 and 13, of oligoclase graphic granite and the compositions here calculated, especially for $\frac{3}{4}$ Ab:$\frac{1}{4}$ An.

For the system *Quartz:Orthoclase* (microcline, with about 72 Or+28 Ab_{+An}):*Albite* (with about 88 Ab or Ab_{+An} and 12 Or) we must have three individualization-fields, with partial eutectics

respectively about 26 Qu:74 orthoclase (72 Or+28 Ab), about 26 Qu:74 Albite (Ab with little An and about 12 Or) and about 40 Or:60 Ab_{+An} (or 42:58). If we leave the inconsiderable

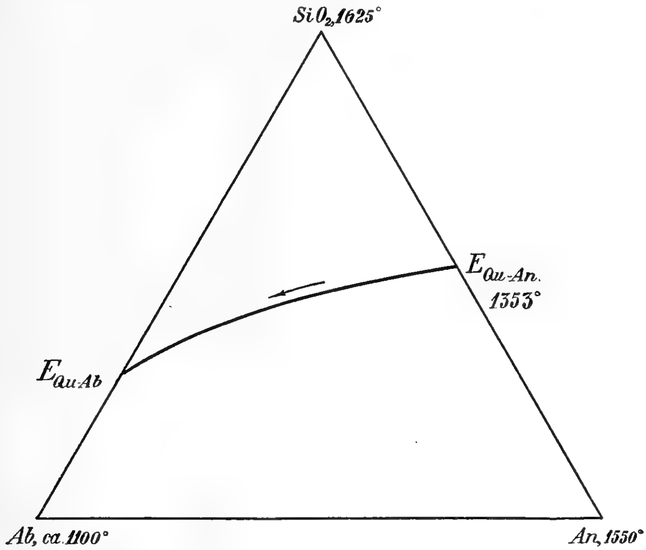

FIG. 5.—The ternary system Ab:An:SiO_2 (horizontal projection)

admixture of An out of consideration, we must suppose for Qu:Or:Ab a ternary eutectic, with about 26 Qu and about 42 Or:58 Ab, consequently about 26 Qu:31 Or:43 Ab (or Ab_{+An}).

We calculate the composition of these eutectics (see *Tscherm. Mitt.*, Vol. XV [1906], p. 385) as shown in Table IV.

TABLE IV

APPROXIMATE CALCULATION OF EUTECTICS QUARTZ: ORTHOCLASE (MICROCLINE): ALBITE (WITH LITTLE AN)

Percentage of	Eutectic Qu:Orthoclase (72 Or:28 Ab+An)			"Ternary Eutectic" Qu:Orthoclase:Albite (with 42 Or:58 Ab+An)				Eutectic Qu:Albite (88 Ab or Ab+An:12 Or)			
Qu	26.0	26.0	26.0	26.0	26.0	26.0	26.0	26.0	26.0	28.0	32.0
Or	53.3	53.3	53.3	31.1	31.1	31.1	31.1	8.9	8.9	8.7	8.6
Ab	20.2	19.2	17.7	42.4	41.4	39.9	36.9	64.6	63.6	60.3	53.4
An	0.5	1.5	3	0.5	1.5	3.0	6.0	0.5	1.5	3.0	6.0
SiO_2	74.7	74.45	74.05	75.55	75.3	74.9	74.15	76.45	76.2	76.45	76.9
Al_2O_3	13.85	14.0	14.25	14.1	14.25	14.55	15.05	14.35	14.5	14.4	14.15
K_2O	9.0	9.0	9.0	5.25	5.25	5.25	5.25	1.5	1.5	1.45	1.45
Na_2O	2.35	2.25	2.1	5.0	4.9	4.7	4.35	7.6	7.5	7.1	6.3
CaO	0.1	0.3	0.6	1.0	0.3	0.6	1.2	0.1	0.3	0.g	1.2

TABLE V

Percentage of	Andesites				Dacite	Spherulit Rock	Dacite	Quartz-porphyry	
	Emmons		(Idd.)	(Lag.)	(Idd.)	(Lag.)	(Lag.)	(Lasp.)	(Streng)
	14a	15a	16a	17a	18a	19a	20a	21a	22a
SiO_2	59.06	60.39	62.00*	62.54	69.36	76.48	75.07	72.24	74.11
Al_2O_3	16.40	16.96	17.84		16.23	12.09	12.15	13.63	13.69
Fe_2O_3	2.88	1.50	......	23.56	0.88	0.95	1.67	......	......
FeO	4.18	3.42	4.40		1.53	......	......	3.05†	1.75
MgO	2.63	3.81	2.64	1.15	1.34	0.39	0.14	0.66	0.05
CaO	4.32	5.41	5.37	4.75	3.17	0.64	0.86	0.95	1.38
Na_2O	5.29	3.37	4.29	3.16	4.06	4.89	4.12	2.95	1.54
K_2O	1.49	2.01	1.47	2.43	3.02	3.78	4.57	5.24	5.67
H_2O, ign.	2.06	2.03	1.66	1.75	0.45	0.77	1.34	1.26	0.56
Total	98.31	99.11	100.13	99.35	100.04	99.96	99.92	100.11	99.90

	Phenocrysts								
	(14b)	(15b)	16b	17b	18b	19b	20b		22b
SiO_2	56.25	55.92	56.41	55.42	(65.77†)	62.14	65.49	......	61.80
Al_2O_3	28.56	28.58	27.39	28.01	21.51	22.30	18.74	......	19.28
Fe_2O_3	0.77	1.00	0.69‡	1.09	tr.	tr.	tr.	......	2.02
MgO	1.06	0.55	0.09	tr.	0.00	tr.	......	......	0.01
CaO	6.54	6.65	9.87	9.12	5.72	3.29	0.39	......	......
Na_2O	5.42	5.66	5.43	5.10	5.92	10.58	3.53	......	0.68
K_2O	0.61	0.66	0.30	0.79	0.83	1.69	9.45	......	12.18
H_2O, ign.	0.33	0.69	......	0.52	0.34	0.20	1.30	......	0.25
Total	99.54	99.71	100.24	100.05	100.09	100.20	98.90	......	100.10

	Glass Basis						Groundmass		
	14c	15c	16c	17c	18c	19c	20c	21c	22c
SiO_2	68.11	68.60	69.94	70.19	76.75	74.59	74.96	74.41	74.44
Al_2O_3	15.56	17.27	15.63		12.32	12.88	13.67	13.39	13.51
Fe_2O_3	0.96	2.09	....	17.19	.....	0.80	1.80	......	......
FeO	2.12	0.58	1.89		1.36	......	......	3.08	2.25
MgO	1.10	0.40	0.28	0.53	0.00	0.30	tr.	0.50	0.01
CaO	2.91	1.72	2.49	2.50	1.18	0.76	0.62	1.38	1.19
Na_2O	3.43	3.30	3.83	3.30	3.55	3.30	2.70	3.27	1.40
K_2O	2.61	4.61	2.85	3.89	3.98	5.35	4.14	4.18	5.31
H_2O, ign.	2.82	2.08	3.25	2.31	0.54	1.03	1.52	1.04	1.34
Total	99.62	100.65	100.16	99.91	99.68	99.01	99.41	101.55	99.45

* +0.17 TiO_2, 0.29 P_2O_5 in the rock.

† In the rock 0.13 and in the groundmass 0.30 per cent MnO.

‡ is FeO.

EXPLANATION

Nos. 14 and 15: H. Emmons, "Island of Capraja, at Elba," *Quart. Jour. London*, 1893. The two feldspars 14*b* and 15*b* isolated by density 2.67—No. 16: A. Hague and J. P. Iddings, "Volcanoes of Northern California, Oregon, and Washington Territory," *Amer. Jour. Sc.*, 3, Vol. XXVI (1883). Hypersthene-andesite.—No. 17: A. Lagorio, "Über die Natur der Glasbasis sowie der Krystallisationsvorgänge in eruptiven Magma," *Tscherm. min. u. petrogr. Mitt*, Vol. VIII (1887); from Hliniker Valley, Hungary.—No. 18: Hague and Iddings (*loc. cit.*); California. The plagioclase is andesine-oligoclase. The determination of SiO_2 in the plagioclase is too high, owing to impurity.—No. 19: Lagorio (*loc cit.*). Spherulitic rock; from Alausi, Ecuador.—No. 20: Lagorio (*loc. cit.*); from Summit County, Colorado. With phenocrysts of quartz and two feldspars, one monocline (analysis No. 20*b*) and the other tricline.—No. 21: Laspeyres (see Zirkel's *Textbook of Petrography*, 1894, Vol. II, p. 177); from Halle, Germany.—No. 22: A. Streng, *Neuer Jahrb. f. Min., Geol. u. Pal.*, 1860, from the Harz Mountains.

In order to establish that the crystallization in the eruptive magmas of the different feldspars and of quartz is in conformity with the physicochemical details which we here have developed essentially on the basis of the analysis of graphic granite, we refer *inter alia* to my earlier statement in *Tscherm. Mitt.*, Vol. XXV (1906).

On pages 340 and 342 we give a small selection (analyses Nos. 14–29) from the numerous analyses, compiled from the literature, partly of porphyritic rocks, with special analyses of (*a*) the whole rock, (*b*) the porphyritic feldspar, and (*c*) the glass or groundmass, and partly of some granites with special analyses of (*b*) the basic concretions (or orbicules), and (*c*) the inclosing rock.

Granites with basic concretions (Nos. 24*b*–26*b*), or basic orbs (Nos. 27*b*–29*b*) are shown in Table VI on p. 342.

We call special attention to the following:

1. In the intermediate and the acid eruptive rocks, which contain the ordinary admixture of ferromagnesian silicates and iron, or titanic iron, ore (especially magnetite and ilmenite), an essential part of these minerals crystallizes at an early stage. A small quantity of Fe_2O_3, FeO, and MgO, however, is left in the remaining magma. This appears in the solidified rocks as the glass basis or groundmass in the porphyritic rocks, or as the intervening mass between basic concretions or orbicules in granites.

In the final, very complicated eutectic, chiefly consisting of feldspars and quartz, there usually seem to appear 0.5 per cent Fe_2O_3, 0.25 per cent FeO, and 0.1–0.25 per cent MgO; the figures,

TABLE VI

Percentage of	Basic Concretions					Basic Orbs			
	23*b*	24*b*₁	25*b*₁	25*b*₂	26*b*	27*b*	28*b*	29*b*₁	29*b*₂
SiO_2	53.80	54.73	56.01	56.53	64.39	55.72	65.57	61.10	68.02
TiO_2	0.77	tr.	1.13	1.40	tr.	0.57	......	0.51	......
Al_2O_3	19.20	14.02	15.19	16.47	15.99	21.35	17.46	15.55	15.31
Fe_2O_3	7.60	2.34	2.34	1.58	1.47	.....	4.15	2.10	0.59
FeO		4.92	4.89	5.40	5.98	8.81		2.21	2.14
MnO	......	......	0.40	0.20	tr.	0.36	......	......	......
MgO	4.80	7.40	4.67	2.67	1.67	0.63	2.53	6.30	3.41
CaO	5.70	10.20	4.85	4.90	2.57	5.10	2.49	1.12	1.53
Na_2O	2.16	2.98	5.66	5.59	4.96	5.71	2.14	1.23	2.85
K_2O	5.08	2.67	2.16	3.80	2.46	1.23	4.23	9.38	5.67
P_2O_5	1.20	tr.	0.53	0.27	tr.	......	......	......	......
Ign.	1.28	1.23	1.26	0.93	0.95	0.46	1.26	1.80	0.46
Total	101.59	100.99	99.21	99.98	100.44	99.94	99.83	100.58	99.98
	Intervening Mass								
	23*c*	24*c*	25*c*₁	25*c*₂	26*c*	27*c*	28*c*	29*c*	
SiO_2	74.40	70.44	71.90	73.69	73.70	70.05	68.27	71.71	......
TiO_2	......	......	0.35	0.28	tr.	0.19	......	......	......
Al_2O_3	13.91	15.63	14.12	12.46	14.44	14.78	15.59	15.05	......
Fe_2O_3	1.39	1.34	1.20	1.21	0.43	.....	2.13	1.11	......
FeO		1.12	0.86	1.75	1.49	3.37		0.29	......
MnO	......	......	0.05	0.15	tr.	0.22	......	......	......
MgO	0.28	0.55	0.33	0.17	tr.	0.44	1.19	0.56	......
CaO	0.61	1.98	1.13	0.36	1.08	3.42	1.93	1.42	......
Na_2O	4.65	4.03	4.52	4.47	4.21	3.10	3.21	3.39	......
K_2O	4.36	5.18	4.81	4.92	4.43	4.13	5.37	5.43	......
P_2O_5	......	......	0.11	0.04	tr.	......	......	......	......
Ign.	0.65	0.55	0.60	0.38	0.61	0.42	1.56	0.61	......
Total	100.25	100.82	100.35	100.09	100.39	100.12	99.25	99.77	......

EXPLANATION

No. 23: See Rosenbusch, *Elemente der Gesteinslehre*, 1901; from Pelvoux.—No. 24: Graber, *Jahrb. d. k.-k. geol. Reichsanstalt*, Wien, 1897; from Topla in Carinthia.—No. 25: Clarke, *U.S. Geol. Survey Bull. 168*, 24; from Mount Ascutney, Vt.—No. 25b_2 and c_2, granite porphyry.—No. 26: J. A. Phillips, *Quart. Jour. London*, 1880; from Peterhead, Scotland.—No. 27: H. Bäckström, *Geol. Fören. Förh.*, Stockholm, 1894; from Kortfors, Sweden.—Nos. 28–29: K. v. Chrustschoff, "Über holokrystalline makrovariolithische Gesteine," *Mém. de l'académie des sc. de St. Pétersbourg*, 1894. No. 28 from Altai; No. 29 from Fonni, Sardinia.

however, especially for MgO, depend somewhat upon the Mg-bearing mineral component in question (biotite, hypersthene, etc.).

2. As previously mentioned and as illustrated by analyses Nos. 14–19 and 20–22, plagioclase crystallizes when there is a surplus of Ab+An; orthoclase, on the contrary, when there is a surplus of Or in the original solution. The boundary lies, as previously explained, at about 0.4 Or:0.6 Ab+An. In the plagioclase which crystallized first, relatively much An appears, consequently relatively much CaO (and Al_2O_3) (cf. the analyses Nos. 14*b*–19*b*). In consequence, the remaining magma shows a decreasing percentage of CaO, and this in the "granite eutectic" falls to 0.25, 0.5, or 1 per cent CaO, or, with predominating plagioclase in the eutectic, not quite so low.

If we leave magmas with only a trifle of Na_2O (Ab), or of K_2O (Or), out of consideration—where the crystallizing orthoclase absorbs practically all of the Ab or the crystallizing albite or albite-oligoclase the Or—the contents of Or (or K_2O) in the magma remnant increases by the crystallization of plagioclase (cf. the analyses Nos. 14*c*–19*c* and 24*c*, 27*c*), and the contents of Ab (or Na_2O) in the magma remnant increases by the crystallization of orthoclase (cf. No. 21*c*). But we especially emphasize that this *relative increase has a limited course*, and that *the limit of about 0.4 Or:0.6 Ab or Ab+An is not exceeded*, or only perhaps now and then somewhat exceeded because of supersaturation. In this matter we especially refer to the analyses of the glass basis or groundmass of porphyritic rocks. In judging these analyses, however, we must take into consideration that the glass, as is shown, for example, by the water content, is always or nearly always somewhat decomposed, whereby especially a little alkali will be extracted. We further refer to the intervening material between basic concretions or orbicules in granites.

As examples these intervening masses from casually chosen granites show:

In Nos. 28 and 29, where orthoclase crystallized first, respectively 0.45 Or:0.55 Ab_{+An} and 0.46 Or:0.54 Ab_{+An}.

In No. 23, where both orthoclase and plagioclase seem to have crystallized at an early stage: 0.37 Or:0.63 Ab_{+An}.

In Nos. 24, $25c^1$ and c^2, 26, and 27, where plagioclase (or perhaps predominant plagioclase and some orthoclase) crystallized at an early stage, respectively: 0.40, 0.38, 0.41, 0.38, and 0.36 Or. The remainder is Ab_{+An}.

These values, calculated from the analyses, for the proportions Or:Ab_{+An} need, however, a small correction, as we have taken for granted that the whole quantity K_2O, Na_2O, and CaO form respectively Or, Ab, and An, while in reality a trifle K_2O (and a still smaller amount of Na_2O) in several cases enters into biotite.

3. The glass basis, or the groundmass, in andesites (with at least about 56 per cent SiO_2, that is to say, with at least so much SiO_2 that a little of the independent component quartz entered into the melted magma) and also in dacites, trachytes, rhyolites, quartz-porphyries, etc. (with max. about 72 per cent SiO_2) without exception shows an increased percentage of SiO_2 as compared with the entire rock. Gradually as the crystallization, for example, of an andesite with 59 per cent SiO_2 (No. 14*a*) advances with solidification of magnetite, ferromagnesian silicates, and medium-basic plagioclase, the temperature drops, and simultaneously the mother liquor becomes richer in SiO_2. By rapid cooling the viscosity increases and causes the cessation of the crystallization; in other words, the fluid remnant stiffens into glass, with varying percentages of SiO_2 according to the time at which the crystallization ceased. And this point of time may lie even considerably lower than the stage of the final eutectic. This we may illustrate by giving the percentages of SiO_2 in the entire rock and in the glass basis (or in some cases the groundmasses); the latter, however, always contains a little H_2O, showing that it is somewhat decomposed. (See my treatise in *Tscherm. Mitt.*, Vol. XXIV [1905].)

ANDESITES

Percentage of SiO_2 in									
Rock	56.8	57.8	58.1	59.1	60.1	60.4	62.0	62.3	62.5
Glass basis	64.5	65.1	70.8	68.1	68.7	68.6	69.9	67.0	70.2

DACITES

Percentage of SiO_2 in						
Rock	65.6	68.3	69.4	65.5	65.3	67.3
Glass basis	71.9	74.8*	76.75	70.2	70.2 and 73.6	72.4

* Goundmass.

Further we include a series of rhyolites, dacites, spherulites, and quartz-porphyries, in which phenocrysts of feldspar (orthoclase or plagioclase) as well as of quartz usually appear. Gl. = glass basis; Grm. = groundmass; Sph. = Spherulite.

TABLE VII

	Percentage of	Dacites		Rhyolites					Spherulitic Rocks		
		Sph.	Grm.	Grm.	Sph. and Gl.	Gl.	Sph. Gl.		Gl.	Sph. Gl.	
Rock	SiO_2	74.6	75.07	72.5	75.8	73.0	71.4		76.5	72.8	
Gl., Grm.,	SiO_2	76.05*	74.96	77.5†	72.7	72.6	74.5	72.5	74.6	73.7	70.7
or Sph...	H_2O	1.2	1.5	1.4	0.9	4.55	1.4	4.4	1.0	1.4	3.7

*Only 10.24 per cent Al_2O_3. Somewhat decomposed.
†Only 11.52 per cent Al_2O_3. Somewhat decomposed.

	Percentage of	Spherulitic Rocks with Spherulites and Glass Basis					Quartz-Porphyries with Groundmass			
Sph........	SiO_2	73.2	74.4	75.4	73.4	Qu-porph...	72.0	72.2	74.1	75.2
Gl.........	SiO_2	72.4	72.7	73.0	73.1	Grm.......	74.0	74.4	74.4	(76.8?)
	H_2O	1.4	1.1	3.6	0.9			1.0	1.3	0.8

The glass basis, or the groundmass, in the porphyritic rocks consequently shows:

a) In rocks with about 60 per cent SiO_2, a sometimes very considerable increase in the percentage of SiO_2 (for example, from 58.1 to 70.8 per cent SiO_2) and in rocks with about 65 or 65–70 per cent SiO_2 a smaller increase, though in undecomposed state not above 75 per cent SiO_2.

b) In rocks with about 73–75 per cent SiO_2 we find, on the other hand, about the same percentage of SiO_2 in the glass basis or groundmass as in the entire rock. The analyses show some slight variations, partly in one and partly in the other direction. But this is certainly caused in some cases only by slight inaccuracies in the relatively old analyses, and in others by the groundmass and especially the glass basis (probably without exception) being a little decomposed, as shown by a little H_2O.

These law-governed relations, which are established by numerous analyses, may depend on the fact that in a magma consisting chiefly of Qu and Or, Ab and An components, with a

surplus of feldspar components, the crystallization of feldspar may continue without a simultaneous secretion of quartz until the eutectic boundary-line between quartz and the feldspar components has been reached. When this has occurred, however, a simultaneous crystallization of feldspar and quartz commences, with only a quite inconsiderable change of the SiO_2 percentage of the magma remnant, while we constantly more and more approach the "ternary" eutectic: Qu:Or:Ab+An (with a trifle magnetite and ferromagnesian silicate).

The groundmass in the quartz porphyries and the closely related rocks consists, as is well known, in some cases of microfelsite and in others of granophyre, and these structural forms indicate a *simultaneous* crystallization of quartz and the feldspar in question. The final crystallization consequently took place also with regard to the structure at a eutectic or eutectic boundary-line.

Especially acid quartz porphyries (with more than 75 per cent SiO_2) show phenocrysts of quartz and feldspar in about equal amounts, but groundmasses of normal microfelsite or granophyre, that is to say, with relatively less quartz than among the phenocrysts. The groundmass, consequently, here must have grown a little more basic than the original rock. I lack material, however, to prove this by chemical analysis.

If we now turn to the deep-seated igneous rocks, we find that the quartz-norites and quartz-gabbros (with about 1 to 5 or 6 per cent quartz), the quartz-diorites, the quartz-syenites, etc., prevailingly show that the quartz first began crystallizing at a relatively late stage. As we shall explain later (Figs. 17 and 18) when treating of the quartz-norites, this crystallization of quartz at a late stage took place, not by itself, but simultaneously with the final crystallization of the feldspar (the plagioclase) and the ferromagnesian silicate in question. In the granite porphyries, which contain but little ferromagnesian silicate and magnetite but are especially rich in feldspar, with about 66–70 per cent SiO_2, the crystallization commenced with the solidification of some feldspar.

The crystallization in ordinary granites usually commenced with a solidification of some magnetite and ferromagnesian silicate,

while the feldspar first commenced crystallizing at a somewhat later stage. The sequence of the *commencement* of the crystallization in the granites is in most cases (1) pyrite, zirkon, apatite, etc.; (2) iron ore and ferromagnesian silicate; (3) feldspar; (4) quartz. But it appears from the structure that the ferromagnesian silicate, especially biotite, continued crystallizing after the commencement of the solidification of both feldspar and quartz, and that the feldspar continued crystallizing also during the segregation of the quartz. In most of the granites, however, we are unable to determine with accuracy from the structure, the quantitative proportions of feldspar and quartz during the intermediate and later stages of the crystallization.

The case is complicated by the fact that the granite magma contains, besides the usual ferromagnesian silicate, Or, Ab, An, and Qu components, some H_2O, probably partly entering into a SiO_2 combination, for example, as H_2SiO_3 (?), and the latter was not split up until a later stage of the crystallization period. If this supposition is correct, the consequence will be a somewhat reduced quantity of the independent quartz component during the first part of the crystallization period—that is to say, during the first part of the crystallization the feldspar was relatively more abundant than that corresponding to the proportion calculated from the relation between feldspar and quartz in the resulting solid rock.

We have an instructive orientation on the composition of granite magmas at a late stage of the solidification in the composition of the intervening masses between basic concretions, or orbicules, in granites, which show these structural elements. (See analyses Nos. 23*c*–29*c*.) These intervening masses prove throughout that during the crystallization a displacement of the composition of the magma remainder took place in the direction of the—in other ways determined—"granite eutectic," and we especially emphasize that the analyses of the intervening masses Nos. 23*c*, $25c_2$, and 26*c* almost exactly correspond with the "granitic eutectic."

Above we have only considered granites with relatively basic concretions, or orbicules. But from strongly *acid* granites, with about 78–80 per cent SiO_2 in the whole rock, we know a couple of

examples[1] of orbicular structure, the orbs chiefly consisting of *quartz*, and in addition some sillimanite and tourmaline (!). The intervening mass between the orbs here contains less SiO_2 (and less quartz) than the orbs, therefore reckoned from the acid pole there here appears a displacement of the residual magma in the direction of the eutectic, quartz:feldspar. In these cases we have, however, the complication that the orbs contain much tourmaline and sillimanite, the latter being very little soluble in acid magma.

Finally we compile a series of analyses of glass bases (Gl.), respectively groundmasses (Grm.), and spherulites (Sph.) from porphyritic rocks, and intervening masses from granites with basic concretions, etc. These analyses represent approximately the composition of the residual magmas resulting from far-advanced

TABLE VIII

ANALYSES OF THE FINAL PRODUCT OF THE CRYSTALLIZATION OF ACID ROCKS—THE GRANITIC EUTECTIC

	No.	SiO_2 Without H_2O	SiO_2	Al_2O_3	Fe_2O_3	FeO	MgO	CaO	Na_2O	K_2O	H_2O	Total
Gl., Sph.	30	73.4	72.69	.....	15.04	....	tr.	0.25	1.75	8.85	0.94	**99.52**
Grm.	31	74.0	74.0	.....	.....	....	.....	.....	0.60	7.57	.....	.....
Glass basis	32	73.4	72.44	.....	16.28	....	0.20	0.59	2.12	6.49	1.35	**99.47**
	33	74.4	73.57	13.80	.....	1.54	0.26	0.99	3.09	5.74	1.08	**100.07**
	34	73.3	72.35	13.97	1.29	....	0.46	0.72	3.58	5.38	1.37	**99.12**
	35	73.7	73.05	14.67	0.89	....	0.26	0.97	3.99	5.11	0.91	**99.85**
	19*c*	75.3	74.59	12.88	0.80	....	0.30	0.76	3.30	5.35	1.03	**99.01**
Intervening mass of granites	25*c*₂	74.0	73.69	12.46	1.21	1.75	0.17	0.36	4.47	4.92	0.38	**100.09**
	23*c*	74.9	74.40	13.91	1.39	....	0.28	0.61	4.65	4.36	0.65	**100.25**
	26*c*	74.1	73.70	14.40	0.43	1.49	tr.	1.08	4.21	4.43	0.61	**100.39**
Spherulites	36	74.0	73.21	12.90	2.10	....	0.27	0.88	4.83	4.75	1.04	**99.98**
	37*c*₁	75.1	74.52	12.97	2.02	....	0.25	0.92	4.26	4.53	0.83	**100.30**
	38	74.7	73.72	12.91	1.37	....	0.25	1.37	4.02	4.45	1.36	**99.45**
Gl.	39	73.5	72.70	13.79	1.01	....	0.65	2.07	4.93	4.33	1.10	**100.48**
Sph.	37*c*₂	74.0	73.42	14.29	1.01	....	0.43	1.00	5.61	3.19	0.84	**99.79**
	40	74.8	74.36	14.46	1.62	....	0.44	1.49	6.11	1.49	0.57	**100.54**

[1] From Kragerö and Modum in Norway and from Pine Lake in Ontario (lecture by W. C. Brögger on the Kragerö locality, in *Kristiania Vidensk. Seisk*, 1901, and Frank D. Adams, *Bull. Geol. Soc. Amer. No. 9*, 1898, see résumé in my treatise in *Tscherm. Mitt.*, Vol. XXV [1906]).

EXPLANATION

No. 30: Spherulitic glass basis from liparites.—No. 31: groundmass from quartz-porphyry.—Nos. 32, 33: Glass basis from trachytes.—Nos. 34, 35, 39: Glass basis from spherulitic rocks—Nos. 36, 37c_1 and c_2, 38, and 40: Spherulites from spherulitic rocks.

Nos. 30, 32, 34, 35, 36–40 reprinted from the above-cited excellent treatise of Lagorio, 1887.—No. 31, see Zirkel's textbook.—No. 33, Williams, *Neues Jahrb. f. Min., Geol. u. Pal.*, Beil., Bd. V, 1887 (see also my treatise in *Tscherm. Mitt.*, Vol. XXIV [1905]).

crystallization. In judging these analyses we must take into consideration that throughout a little alkali was probably extracted from the glass basis, so that the determined percentages of alkalies may be a trifle too low. The analyses are generally arranged according to decreasing K_2O (Or) or increasing Na_2O (or Ab+An).

We especially direct attention to the close accordance between these analyses of the residual magmas resulting from the solidification—partly from dike and surface rocks, and partly from deep-seated rocks[1]—and the eutectic Qu:Or:Ab+An, calculated on the basis of the graphic-granite analyses and the theoretical explanations. (See the table, p. 339.) The accordance is especially pronounced when we take into consideration that the percentage of SiO_2 in the analysis of graphic granite will be reduced about 1 per cent when 1 or 2 per cent of ferromagnesian silicate and magnetite is added, and that the analyses of groundmasses, etc., which are rich in plagioclase, contain a little more CaO (or An) than the graphic granites calculated in the table, page 339, where only a small admixture of An is presupposed. The analyses in the table, page 348, of the residual magma represent the *granitic eutectic*, consisting of predominant Qu, Or, and Ab+An, with the addition of quite small admixtures of iron oxide (Fe_3O_4) and ferromagnesian silicate.

If we leave the latter quite subordinate admixtures out of consideration, the analyses Nos. 25c_2, 23c, 26c, 36, 37c, and 38 almost exactly represent the "ternary" eutectic Qu:Or:Ab (or Ab+An) with nearly exactly 0.4 Or:0.6 Ab (or Ab+An). And this we may by a short catchword name the "*ternary*" *granitic*

[1] Regarding the inconsiderable influence of the pressure on the composition of the eutectic we refer to a following chapter.

eutectic. This term strikes the essential point, since the eutectic in question consists practically only of Qu, Or, and Ab. To this must, however, be added a quite small admixture of An, iron oxide, and ferromagnesian silicate, so that the eutectic in reality is more complicated. In order to avoid misunderstanding I have therefore put the term "ternary" in quotation marks.

[*To be continued*]

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INTRODUCTION

The nature of the northeastern termination of the great overthrust block of the earth's crust, bounded on three sides by the Pine Mountain fault, the Hunter Valley fault, and the Jacksboro cross fault of Tennessee, has long been an unsolved problem to students of Appalachian structural geology. Many geologists have noted the rather abrupt ending, near the breaks of Big Sandy River, of the imposing barrier of Pine Mountain and its replacement to the northeastward by the irregular ridges and valleys of the unbroken coal field, but the manner in which the great anticlinal fold and the resulting thrust fault died out has not until recently been satisfactorily solved.

In 1916 Hinds,[2] in his report on the Clintwood and Bucu quadrangles, called attention to a zone of disturbed rocks nearly at right angles to the general lines of disturbance in this region and extending partly across the trough of coal-measure rocks from Big A Mountain to Skegg Gap on Pine Mountain. Hinds attributed the disturbance in this zone to the same forces that produced the Hunter Valley fault on the southeast and the Pine Mountain fault on the northwest, but he failed to perceive its significance, for he thought it was limited to certain areas and did not extend entirely across the synclinal block.

In April, 1920, Mr. M. R. Campbell, in charge of geologic work in this coal field for the United States Geological Survey, called attention to the possibility of the belt of disturbed rocks mapped

[1] Published by permission of the Directors of the U.S. Geological Survey and the Virginia Geological Survey. The illustrations were prepared for the Virginia Geological Survey.

[2] Henry Hinds, "The Coal Resources of the Clintwood and Bucu Quadrangles, Virginia," *Virginia Geol. Survey Bull. 12* (1916).

by Hinds being but part of a continuous fault or zone of faulting from the Hunter Valley fault at Big A Mountain to Skegg Gap in Pine Mountain, and the author was requested to examine the region as carefully as the limited time at his disposal would permit, in order to establish the character and extent of the movements that produced the disturbance. The result of his examination was the establishment of the presence of an overthrust fault entirely across the great crustal block, thus showing that it is bounded on all four sides by overthrust faults and that it has moved bodily to the northwest a distance of many miles. The results of his studies and their application to the mechanics of the problem of the overthrusting of this great mass of strata for at least six miles are here set forth.

The fault bounding the crustal block on the northeast, which, on account of its general agreement with the course of Russell Fork, is here called the Russell Fork fault, was mapped in connection with coal investigations carried on co-operatively by the Virginia Geological Survey and the federal Geological Survey. The areas mapped as undifferentiated buckled and faulted rocks by Hinds,[1] on the Clintwood and Bucu quadrangles, were subjected to careful study by the writer to determine whether or not there was a continuous break across the coal-measures trough from the vicinity of Big A Mountain to Skegg Gap, but in the two weeks spent on this study there was not time to cover much of the area lying on either side of this zone and the structure contour maps of the report by Hinds furnished many data in compiling the sections shown below and in deducing the amount of displacement.

The writer is indebted to Mr. M. R. Campbell for many helpful suggestions and much assistance in the course of the study.

Hinds,[2] in his report on the coal resources of the Clintwood and Bucu quadrangles, describes the structure of the northeastern end of the Middlesboro syncline in considerable detail. His studies here and farther northeast in Buchanan County[3] have shown that

[1] *Op. cit.*

[2] *Op. cit.*

[3] Henry Hinds, "Geology and Coal Resources of Buchanan County, Virginia," *Virginia Geol. Survey Bull. 18* (1918).

the great overthrust of Pine Mountain suddenly becomes very much less severe at Skegg Gap, and from there northeastward the structure is essentially a low anticline broken by a minor overthrust which decreases rapidly in extent of thrust and comes to an end a few miles into Buchanan County. He considered that the principal Pine Mountain overthrust was cut off at the northeast end by the Skegg Gap fault which he mapped as far as Russell Fork. Between this point and Big A Mountain he has mapped a number of narrow areas of faulted and buckled rocks which he describes in some detail and in explanation of which he postulates lateral shearing with the southwest side moving northward with some overthrusting and buckling against the northeast side. He states that succeeding this movement there was normal faulting along this line in which the southwest side was downthrown. His evidence for this belief is not clear, and his several areas of disturbed rocks are separated by areas in which he found no evidence of movement.

DESCRIPTION OF THE CUMBERLAND BLOCK[1]

The structure of the area concerned in this paper has been described in considerable detail at many points by previous writers.[2] It is not the writer's purpose to give here a thorough description of the structure or topography but rather to point out briefly their alient features.

The "remarkable quadrilateral block" whose southwestern extremity was first recognized by Safford and described in detail by Keith extends from the valley of Cove Creek in Campbell County, northeastern Tennessee, northeastward for one hundred and twenty-five miles to the valley of Russell Fork of Big Sandy River in Dickinson and Buchanan counties, Virginia. It is surprisingly uniform in width, averaging about twenty-five miles, is bounded on the northwest by the Pine Mountain fault and on the southeast by the Hunter Valley fault and the closely associated Wallen Valley fault, which is developed only from near Big Stone

[1] This name is here applied for the first time.

[2] J. M. Safford, *Geology of Tennessee* (1869); M. R. Campbell, *Geologic Folios 12 and 59*, Arthur Keith, *Geologic Folios 33 and 75*, G. H. Ashley and L. C. Glenn, *Prof. Paper 49*, all of the *U.S. Geol. Survey*.

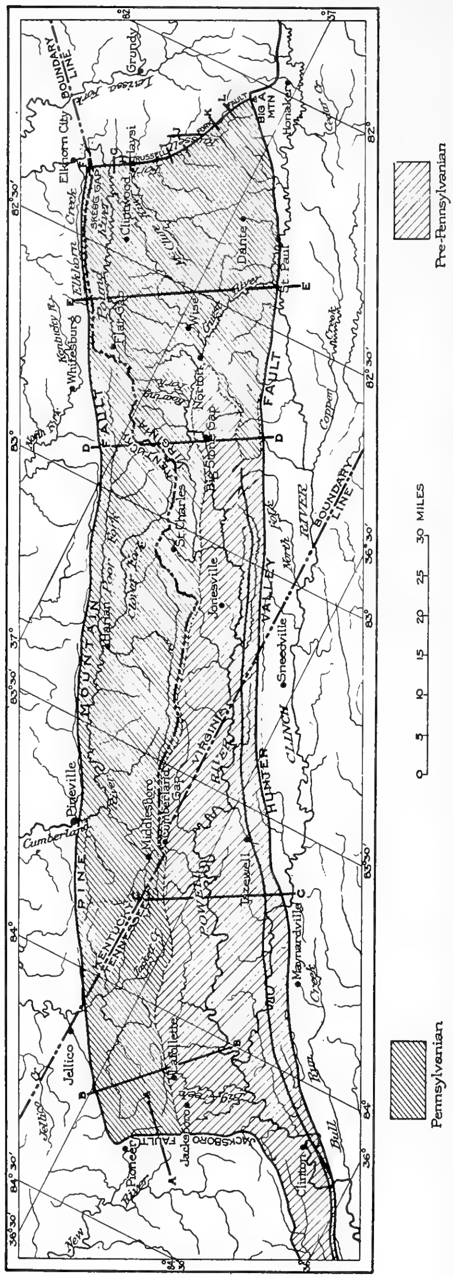

FIG. 1.—Areal map of the Cumberland block showing its location in the states of Virginia, Kentucky, and Tennessee and locations of the structure sections of Figures 3 and 4.

Gap southwestward. The southwest end is terminated by the Jacksboro cross fault and the northeast end by the Russell Fork cross fault to be described below in more detail. The general relations of these boundary faults may be more clearly seen by reference to the map and diagram, Figures 1 and 2, and to the structure sections, Figures 3 and 4.

From Norton, Virginia, northeastward, coal-measure rocks are exposed at the surface throughout the entire width of the block; but from Norton southwestward the block may be divided into two parts, that part lying northwest of Stone and Cumberland mountains being synclinal in structure and composed of coal-measure rocks, whereas that part lying southeast of these mountains is anticlinal in structure and composed of rocks of very much greater age. The syncline, which is now generally known as the

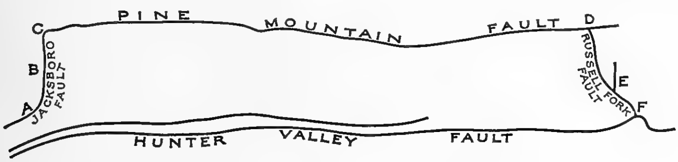

FIG. 2.—Outline diagram of Cumberland block showing the bounding faults

Middlesboro syncline, is a broad, flat-bottomed trough at the northeast end of the block, but farther west in the vicinity of Dante an arch appears which develops rapidly westward into the Powell Valley anticline that constitutes the southern part of the block. Both the Middlesboro syncline and the Powell Valley anticline are characterized by steep dips on the northwest limbs and gentle dips on the southeast limbs, as shown in the sections. The erosion by Powell River and its tributaries of the rising crown of the Powell Valley anticline accounts for the exposures of pre-Carboniferous rocks in the southern portion of the block and the narrowing of the coal-measure portion on the north. The general relation of these different structural units and their expression in the areal relationships of the Pennsylvanian and pre-Pennsylvanian rocks may be seen by reference to Figure 1.

The topography of the block is intimately related to the structure. Pine Mountain throughout its entire length is a conspicuous

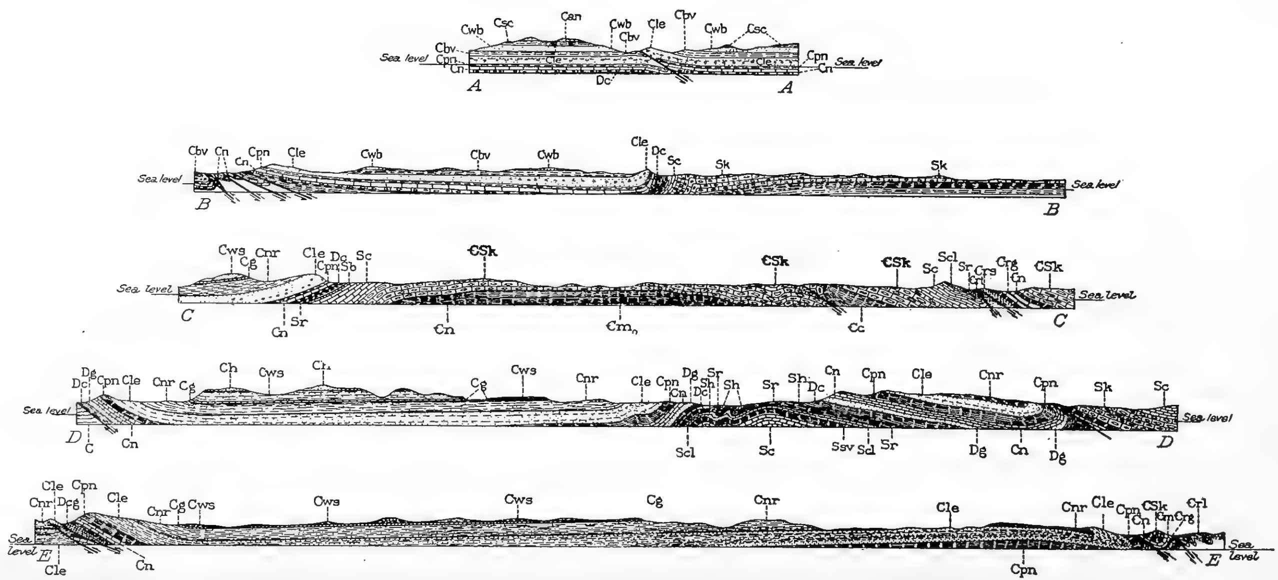

FIG. 3.—Sections showing general structure of the Cumberland block. Sec. *A–A* is seen as looking northwest, the others as looking northeast. For locations see Figure 1. (These sections have been compiled and somewhat generalized from *Geologic Folios 12, 33, 59, and 75*, *Professional Paper 49*, and *Bulletin 541 f.*, all of the U.S. Geol. Survey.)

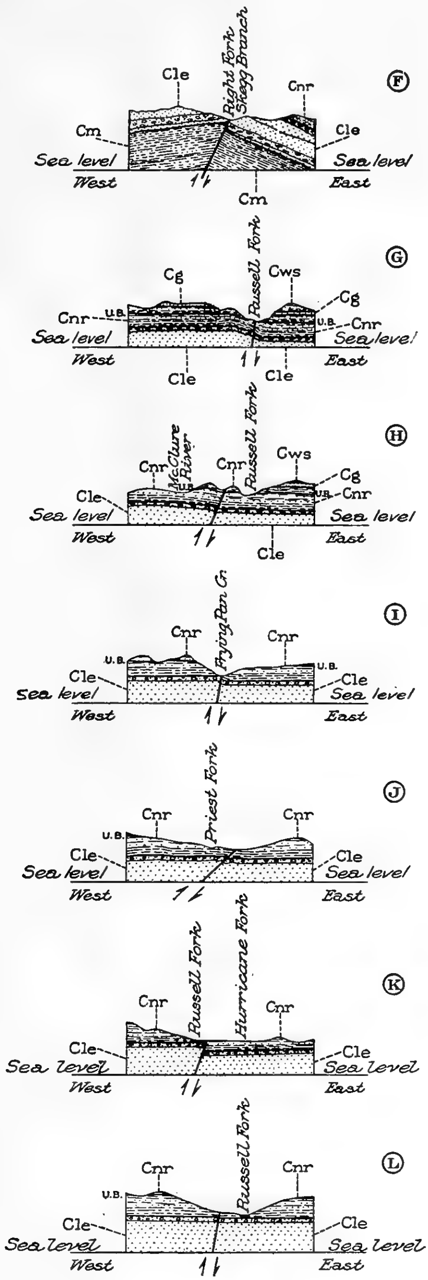

FIG. 4.—Sections showing the character of the Russell Fork fault at several places. For locations see Figure 1. Sections are shown as looking north.

barrier especially on its northwest face, where at many points it rises in less than a mile one thousand to two thousand feet above the streams which parallel it. For nearly ninety miles no stream crosses it and in the entire distance of one hundred and twenty-five miles not over half a dozen roads afford passage from one side to the other. Its crest is the resistant conglomerate of the Lee formation, which marks the edge of the overthrust block. Cumberland Mountain and Stone Mountain are composed of the same formation, which is steeply upturned in that part of the fold common to the syncline and anticline mentioned above. Black Mountain and parts of Sandy Ridge are residual mountains left in the dissection of the nearly horizontal coal measures. Big A Mountain, according to Hinds,[1] is composed of resistant sandstones of the Rockwood formation, overthrust on the coal measures.

The surface features of that portion of the block within the coal field are those of a maturely dissected plateau with sharp-crested ridges and V-shaped valleys, for the most part without valley flats. The surface of the pre-Carboniferous portion is rolling, with sink holes in the limestone portion and some, though not at this particular point very striking, allineation of ridges and valleys with the northeast-southwest trend of the Appalachian structure. The surface configuration of the area and the control of topography by the great structural features is admirably shown on the contour maps of the United States Geological Survey, to which the reader is referred for further detail.

THE RUSSELL FORK FAULT

The Russell Fork fault differs from the faults which bound the Cumberland block on the other three sides in that it is not a low-angle overthrust and that in it the greatest displacement is in a horizontal direction with comparatively little vertical movement. Its trace[2] is closely followed except at a few places by Russell

[1] Henry Hinds, "The Geology and Coal Resources of Buchanan County, Virginia," *Virginia Geol. Survey Bull. 18* (1918), pp. 58–59.

[2] "Trace" of a fault is here used in its mathematical sense of the line of intersection of one surface with another, i.e., the intersection of the fault plane with the surface of the earth.

Fork of Big Sandy River, and even at those places it is marked by the allineation of minor drainage lines or surface features which would not otherwise be easily explained.

Erosion of crushed and weakened rocks along the fault trace has produced the low saddle at Skegg Gap and the saddle in the point of the spur west of Russell Fork and one mile north of B.M. 1221[1] on the Clintwood quadrangle.

From a point one-half mile upstream from B.M. 1282 to B.M. 1221 Russell Fork flows in a course somewhat farther northeastward than in adjacent parts of its course up and down stream. In the high land which lies southwest of this part of the river and northwest of the village of Haysi are cut two short cleftlike hollows which are closely aligned with the fault trace as located to the north and south, and have without doubt been determined by the presence of the weaker rock in the zone of deformation adjacent to the fault. One of these hollows enters the river valley just at the railroad bridge east of B.M. 1221 and the other extends from near Haysi south and just to the west of B.M. 1380. These hollows are somewhat straighter and narrower than most of the ravines of similar size which erosion has cut in the rocks of this region, but their most distinctive characteristic is their location where they cut off in part the narrow strip of high land between them and the river. Taken together and with the other topographic features which show alignment, they are very significant, and their locations are not to be explained as accidental.

Between McClure River and Russell Fork, at the close approach before they join, a low saddle in the spur owes its position to the weakness of the rocks along the fault line. The very straight course of Fryingpan Creek from elevation 1,311 feet to its mouth is determined by the fault, and it is interesting to note that this creek has a very slight fall in this part of its course and its bed is graded for the entire distance with ripple-marked sand. Russell Fork leaves the fault trace at a number of points and because of its cutting across the undisturbed and more resistant rocks at

[1] The area crossed by Russell Fork fault is shown in detail on the Regina, Ky., and the Clintwood and Bucu, Va., sheets of the *Topographic Atlas of the United States*. Frequent reference is made to points on these maps in locating the features described.

these points is not so perfectly graded. But the weakness of the disturbed rocks close to the fault has apparently enabled Fryingpan Creek, though a comparatively small stream, to grade its lower course to the temporary base as determined by the rocks over which Russell Fork flows.

The main line of the fault passes somewhat to the north of Abners Gap; from elevation 1,424 feet southeast to the mouth of Carroll Presley Branch it follows Russell Fork for most of the distance, and thence southeast to the point where it is truncated by the main overthrust fault; in the north face of Big A Mountain its trace lies somewhat to the north of the channel of Russell Fork.

A branch leaves the main fault at elevation 1,424 feet, and extends northwestward along the course of Russell Fork to a point in Little Pawpaw Valley about a mile north of Cannady Post Office, and is here named the Little Pawpaw fault.

Along most of its course the rocks northeast of the fault are horizontal or nearly so and undisturbed. The fault plane, or, better perhaps, the planes of movement, for the most part dip at high angles, 75° to 90° to the southwest. That there has been intense compression is shown by the mashed condition of the shale and jointed condition of sandstone on the southwest side of the fault. At numerous exposures in the zone of faulting, slickensides indicate considerable vertical movement which has resulted in lifting the beds on the southwest above those on the northeast, displacing the coal beds by from 50 to 200 feet. Because of the shearing which has brought anticlines into contact with synclines and vice versa, it is difficult to determine the true amount of differential vertical movement, but the essential point is that the vertical movement is slight and that the hanging wall has moved up as a result of thrust. At many points there are planes other than those which bear the vertical slickensides, a series of horizontal slickensides trending closely in the direction of the fault, and usually these surfaces are rubbed and planed much smoother and more nearly plane than the others, indicating, it seems to the writer, that these surfaces are the result of more extensive movement along the fault line than the other planes along which

a slighter movement has taken place, to accommodate the thrust in a direction lateral to the main Russell Fork fault line (Figs. 5 and 6.)

At a few points, notably at Skegg Gap (Fig. 7), the slickensides and planes of movement within the zone of faulting indicate a combination of the main southeast-to-northwest thrust with the side or southwest-to-northeast thrust, making the direction of movement a resultant of the two. The slickensided surface here shows the result of pronounced movement and the white quartz pebbles which are so numerous in the Lee formation are planed off flush

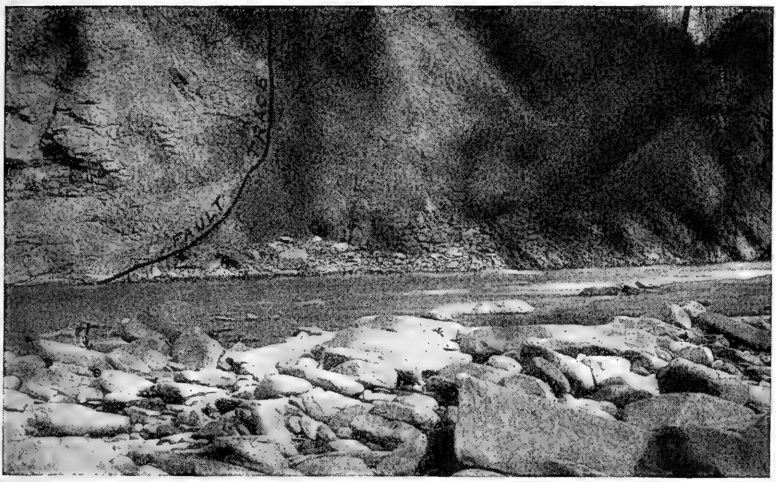

FIG. 5.—View of river bank looking southeast near B.M. 1221. Massive and undeformed sandstone on left of fault trace with deformed shale at right. Strong horizontal slickensides are formed on the face of the sandstone at this place.

with the matrix. It is important to mention that a short distance northwest of Abners Gap the writer saw the most abundant evidence of thrust in the mashing and crushing of shale in a fine exposure and that here the slickensides indicate movement at an angle of 45° to the horizontal in a due north direction. At this point the fault trace is more nearly athwart the main direction of thrust, and here, if anywhere, would be expected evidence of strong overthrust.

In the Little Pawpaw fault there was seen little indication of shearing but abundant evidence of compression and slight overthrusting.

MECHANICS

The history of the deformation (Fig. 8) is conceived to be as follows:

The rocks of the Cumberland block were subjected to strong lateral compression applied from the southeast. The thicker sedimentary rocks west of *A* (Fig. 2, p. 355) seem not to have yielded as did those to the east, and acted as a buttress against which the rocks to the east were deformed. The compressional stress was much more intense at the Tennessee end of the block and the first result of the stress was the folding of the Powell Valley anticline and of the lateral anticline which later broke and formed the Jacksboro cross fault. Between *A* and *B* Keith[1] found evidence of this now broken anticline which was the result of deformation of the rocks of the Cumberland block against the more competent buttress on the west.

After the Powell Valley anticline had been in large measure formed and the Jacksboro cross anticline had probably reached its full development, the stresses were then transmitted across the block, and yielding farther northwest resulted in the folding of the rocks into the Pine Mountain anticline. It is probable that by this time overthrusting and shearing to the northwest had commenced at the southern end of the Jacksboro cross anticline, for the movement of the rocks of the Powell Valley anticline northwestward differentially with respect to the nearly undisturbed rocks on the west had already been very considerable. With the continued crumpling of the Pine Mountain anticline, the Jacksboro cross fault developed progressively toward the northwest, and, when it reached the then position of the corner *C* of the block, initiated the great Pine Mountain fault.

There had by this time been considerable skewing of the entire block which was pivoted at or near its north corner, with the result that the corner of the block at *A* had been thrust more extensively on the rocks to the west than had the corner at *C*. The overthrust to the west in the Jacksboro cross fault is, however, believed to be only the smaller movement incidental to the skewing of the block, while the main movement in this fault was

[1] Arthur Keith, *U.S. Geol. Survey Geol. Atlas, Briceville Folio No. 33* (1896).

the shearing by which the block to the east moved several miles to the northwest. The writer does not believe that there was in the original stress any distinct southwesterly component, but

FIG. 6.—Slickensided shale near mouth of Lick Creek south of Birchleaf

considers the Jacksboro thrust to be solely the result of the twisting of the block.

The Pine Mountain fault is compound at *C*, consisting of four distinct overthrusts, but becomes more simple northeastward. The faulting which commenced at the southwest developed progressively toward *D* as the stress continued, but naturally the

total displacement was less at *D* than at *C*. It seems likely that there was some slight displacement beyond Skegg Gap and into Dickinson County along the Pine Mountain fault before it was intersected by the Russell Fork cross fault.

In following chronologically the development of the Jacksboro–Pine Mountain line of faulting the Russell Fork cross fault was temporarily omitted. The history of its development is correlated with the events described above, as follows:

Only after there had been considerable development of the Pine Mountain anticline at *C*, and some shearing along the Jacksboro cross fault, was the skewing of the Cumberland block felt at the northeast end. Its first expression was the development of a tension or normal fault starting at *F* (Fig. 2) and extending toward *E* with continued twisting of the block.

The presence of the Little Pawpaw fault, which appears to be primarily the result of such tension incident to twisting, leads the writer to believe that the point, or, perhaps more correctly, the area of pivoting, was somewhat to the north of *E*. On the other hand, evidence of somewhat more pronounced compression along the fault from *D*, part of the distance toward *E*, seems to indicate that compression was even at first dominant in that part of the line. It seems therefore probable that the region of pivoting is located between *D* and *E* but somewhat nearer the latter.

After the extension of the Pine Mountain fault beyond Skegg Gap and the extension of the Russell Fork cross fault beyond *E* as a normal fault, the accumulation at the northeast end of the block of the northwestward-trending stresses, which had long been operative at the Tennessee end of the block, reached the critical point, and the northeast end was broken loose along the line largely determined by the pre-existing normal fault. The line of this break intersected the Pine Mountain fault at Skegg Gap, stopped farther movement in that fault east of that line, and permitted the Cumberland block to be thrust not over two miles northwestward at this end. Since the Russell Fork fault line forms an angle of over 90° with the line of the Pine Mountain overthrust, the overthrusting of the east end of the block brought about compression along the whole extent of the Russell Fork

fault, reversing the condition of tension which produced the normal fault, and producing overthrusting and considerable crumpling of the shale and crushing and jointing of the sandstone adjacent to the fault plane. The net amount of overthrust is very slight, probably at no point reaching 500 feet, and the rocks on the southwest side of the fault are nowhere over 250 feet above those on the northeast. The shearing loose of the northeast end of the block, its overthrust along the already established Pine Mountain

FIG. 7.—View of Skegg Gap looking north along fault line. At this point the resistant basal conglomerate of the Lee on the overridden side (right) is adjacent to the weak Pennington rocks of the overthrust side (left).

fault, and the compression and slight thrusting along the Russell Fork cross fault were the closing events in the history of the Cumberland block as a unit.

There are many especial features which are particularly in accord with this interpretation of the movement of the Cumberland block as a progressive skew with final release of the east end. The skewing of the southwest end of the block first with the Skegg Gap corner remaining longest in place explains admirably the otherwise anomalous facts of rather strong overthrust in the Jacksboro fault, the trace of which is nearly at right angles to

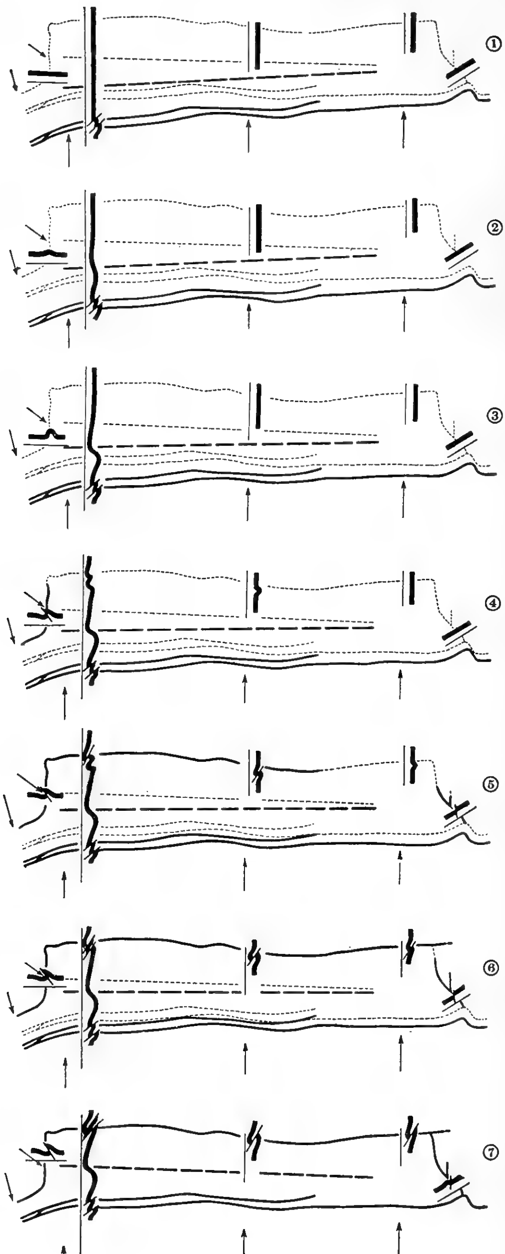

FIG. 8

Schematic structure lines of sections.

Present fault lines placed with reference to the block.

Reference line of block.

Displacement lines of the sections.

Sea-level lines of the sections.

Lines showing present position of fault boundaries and reference line of block.

FIG. 8.—Serial diagram showing history of the deformation and displacement of the Cumberland block as interpreted by the writer. In sketch No. 7 of the series is shown the present condition in outline. The heaviest lines show the structure in the sections schematically. The lightest lines are the "sea-level" lines of the sections and the fault lines of the sections still referring to sketch 7 of the series above. The solid lines of medium weight are the present traces of the various faults. The medium-weight dotted line is an arbitrary axis line the position of which is the same with reference to the block in each sketch of the series.

In sketch 1 of the foregoing series the present fault traces are shown by the light dotted lines of the background and the present position of the axis line by the light and straight dotted line. The then position of the fault traces and of the axis line of the block is shown in sketch 1 by the medium-weight solid and dotted lines respectively. The light "sea-level" lines and fault lines of the sections and the very heavy lines of the schematic sections are the same as in sketch 7. The arrows which point upward in the sketches indicate the direction of the main thrust; the arrows pointing down and to the right at the left-hand end of each sketch indicate the resisting stress of the buttress southwest of Jacksboro as described in the text.

The first sketch assumes the prior formation of the Hunter Valley. The second shows some slight buckling of the southwest end of the block. The third shows more intense buckling here. The fourth shows more intense buckling and extension of the folding eastward along the line of the Pine Mountain fault. In No. 4 also is shown the beginning of the Jacksboro overthrust fault. In No. 5 this fault has extended far around the north side of the block and the Russell Fork fault has been initiated as a normal fault. Sketch 6 shows further extension of the faulting and only a small corner of the block near the northeast corner remains attached. In sketch 7 this small attachment is broken and the block has been thrust into its present position. The relative movement of the block at different stages has been shown by the gradual migration of the heavy dashed axial line toward its final position as shown by the light dotted axial line of each sketch.

the trace of the Pine Mountain fault and of only slight overthrust and more restricted compression in the Russell Fork fault with its trace at a much greater angle with the main overthrust.

The stronger development of the Powell Valley anticline and the presence of the Wallen Valley fault only at the southwest are also strongly in accord with this suggested interpretation.

The greater intensity of stress implied by the compound character of the Pine Mountain overthrust at *C* (Fig. 2) and the probable development of the Powell Valley anticline before the rocks of the block were competent to transmit the stresses to the Pine Mountain fold which later broke in a fault, point strongly to the initiation of the faulting at the south end of the Jacksboro cross fault to allow the necessary shortening of the strata on the northeast side of that fault. The four faults as interpreted by Keith, and corroborated by the displacement of the north limb of Powell Valley anticline in the Briceville quadrangle, give clear evidence of a movement of at least ten miles to the northwest. At Skegg Gap the evidence does not indicate over two miles of overthrust at the most.

In the course of his meditation on this study the writer has made very briefly a few computations, based on extremely general and only very approximate assumptions, which are given below. Their value is solely to indicate orders of magnitude, and it is hoped that they may serve, as they did in the case of the writer, to visualize the immensity of forces involved.

FORCE TO SHEAR AND FORCE TO THRUST

Assumptions

Block 125 miles $\times$ 25 miles $\times \frac{1}{2}$ mile
Density 170 lbs. per cubic foot
Coefficient of friction, mean between rough and smooth granite, 0.60
Shearing strength 200 pounds per square inch
Average extent of overthrust, 6 miles

Results

Force to shear block loose over entire area $= 25 \times 10^{14}$ pounds
Force to move block against friction on horizontal plane $= 23 \times 10^{15}$ pounds
Work done in moving block 6 miles at angle of 5 degrees $= 85 \times 10^{19}$ foot pounds
Equivalent to 420,000 horse-power working for 100,000 years
Estimated coal in block $= 50 \times 10^{9}$ tons

Burning of this coal would produce power enough to move the block 2.2 feet, assuming the usual engine efficiency. It has actually moved an average of at least six miles.

It is especially interesting to note that the force required to shear the block loose over the whole area is only about one-tenth of that required to produce motion against the resistance of friction. Since both forces are proportional to area and only one—that of motion against friction—proportional to thickness, we find that for a block of any area and of a thickness of 287 feet, according to the conditions assumed, the shearing force is just equaled by the force to overcome friction, and as thickness is greater than this amount the latter force is greater in proportion. It is evident, then, that in the case of most overthrust faults the motion of the rock involved against the resistance of friction is more impressive than the production of the break which separated it.

STUDIES OF THE CYCLE OF GLACIATION

WILLIAM HERBERT HOBBS[1]
University of Michigan

I. THE CYCLE OF MOUNTAIN GLACIATION WITHIN MODERATE LATITUDES

In a general discussion of glacial sculpture in mountains,[2] the writer has made use of the terms *grooved* or *channeled upland* and *fretted upland* to describe respectively the early and the late effects of the erosional action of mountain glaciers. The Bighorn Mountains in Wyoming and the Swiss Alps were chosen as type examples of these contrasted erosion surfaces, the characteristics of which are, that in the former large areas of the preglacial upland still remain (Figs. 1, 2, and 3), while in the latter its complete dissection by cirque recession and enlargement has resulted in a system of main and secondary rock palisades described as comb ridges. Between these contrasted land surfaces many gradations exist, though examples of the former are relatively uncommon. Similar to the channeled upland of the Bighorn Mountains, though with less of the preglacial surface retained, are portions of the Uinta and Wasatch mountains, of which excellent illustrations have been supplied by Atwood.[3] All the best examples are furnished by the Rocky Mountains in the interior of the American continent, where the moist westerly winds have been robbed of their moisture in crossing the high Sierra Nevada and Cascade ranges.

In a visit to the Glacier National Park, the writer was impressed with the fact that a type of topography is there represented which indicates a still later stage of sculpture by mountain glaciers than

[1] Illustrations from photographs by the author.

[2] "The Cycle of Mountain Glaciation," *Geogr. Jour.*, Vol. XXXV (1910), pp. 147–53, Figs. 1–19. Also, "Characteristics of Existing Glaciers," pp. 25–40, Pls. 3–9.

[3] "Glaciation of the Uinta and Wasatch Mountains," *U.S. Geol. Survey, Prof. Paper 61* (1909), maps of Pl. 8A.

does the fretted upland as exemplified by the Alps. The most striking peculiarities of this type are found in the unusual number of isolated sharp peaks of monumented aspect (Figs. 4 and 5), and this is combined with a general absence of the comb ridge (a rare example is shown in Fig. 6) and a frequency of unusually low cols or passes (Fig. 7). Unlike the true horns of the Matterhorn type, which in the fretted upland are relatively few in number and may perhaps represent by their summits points near the original surface of the upland, the monuments of the northern Rocky Mountains show a tendency to appear in pairs, and in many

FIG. 1.—View of Mt. Mathews, Bighorn Range, taken from the southeast and showing the character of the preglacial surface. At the left in middle distance is a cirque.

instances at least they are remnants of lower portions of the preglacial surface (Figs. 5, 8).

Both in the Bighorn Range and in the Glacier National Park the glaciers have today nearly or quite disappeared, being now represented by small horseshoe or cliff glacierets only. The earlier conditions of nourishment were, however, as we know from more or less extended studies, notably different from those of today. In the Bighorn Range the glaciers of Pleistocene time extended far down the valleys, where strong terminal moraines are found to mark the limits of their advance.[1]

[1] R. G. Salisbury, "Cloud Peak–Sheridan Folio," *U.S. Geol. Survey;* also, N. H. Darton, *U.S. Geol. Survey, Prof. Paper 51*, pp. 71–91, Pls. 37–36.

FIG. 2.—View of cirque above Seven Brothers Lakes, seen from the ridge south of Trail Lodge, Bighorn Range.

FIG. 3.—Nearer view of the cirque shown in Figure 2. Characteristic surface of preglacial area in foreground.

In the Glacier Park district the Pleistocene glaciers occupied the entire valleys within the range and spread out eastward their aprons of Piedmont type. They also extended westward a long distance down the valley of the Flathead River.[1]

It is here proposed to use the term *monumented* upland to describe the extreme type of mountain sculpture which is represented in the Glacier National Park and which is believed to be due to continued glacial action upon a fretted upland like that of the Alps. Cirque enlargement carried to this stage has sapped the

FIG. 4.—View of Reynolds Mountain, a characteristic monument of the Glacier National Park region. View taken from the trail to Piegan Pass.

main comb ridge so as to largely obliterate the *aiguille* type of crest or *arête* (Fig. 6 shows one of the remaining comb ridges). Matterhorns have in the process been reduced in size as the cols are progressively lowered and widened and are transformed into *arêtes*. The last remnants of the upland to be removed by this continued cirque enlargement are found away from the original divide and outward toward the flanks of the upland, for the reason that in their later stages cirques enlarge excessively on their lateral walls. A good illustration of this tendency is supplied by the

[1] Wm. C. Alden, "Pre-Wisconsin Glacial Drift in the Region of Glacier National Park," *Bull. Geol. Soc. Am.*, Vol. XXIII (1912), Pl. 37. See also by same author, "Glaciers of Glacier National Park," and especially the map opposite p. 32.

FIG. 5.—A pair of monuments of monumented upland seen from the Piegan Trail, Glacier National Park.

FIG. 6.—View of a comb ridge looking across the Piegan Pass, Glacier National Park.

gently sloping summit plane of Quadrant Mountain in the Yellowstone National Park at an elevation of between 9,000 and 10,000 feet[1] (Fig. 9), since Antler Peak and Bannock Peak guard the entrance to the cirque.

It is especially because the comb ridges in the highest levels are precipitous and correspondingly thin that a continuation of

FIG. 7.—Gunsight Pass, seen from Gunsight Chalets looking across Gunsight Lake, Glacier National Park.

the process removes their pinnacles while the broader ridges somewhat farther out and just below the mother-cirques are being sharpened into peaks, both alike through sapping from the cirques.

To bring together the extremes of mountain glacier erosion which are represented by the Bighorn Range and the Glacier National Park with the intermediate stages which connect them, the four generalized plans of Figure 10 have been prepared. In order, these are:

I. The youthful channeled or grooved upland
II. The adolescent early fretted upland
III. The fretted upland of full maturity
IV. The monumented upland of old age

[1] "The Cycle of Mountain Glaciation," *Geogr. Jour.*, Vol. XXXVI (1910), Figs. 8, 14.

These four stages are perhaps best illustrated by the Bighorn Range, the mass of Snowden in the Welsh highland, the Alps, and Glacier National Park (Fig. 10).

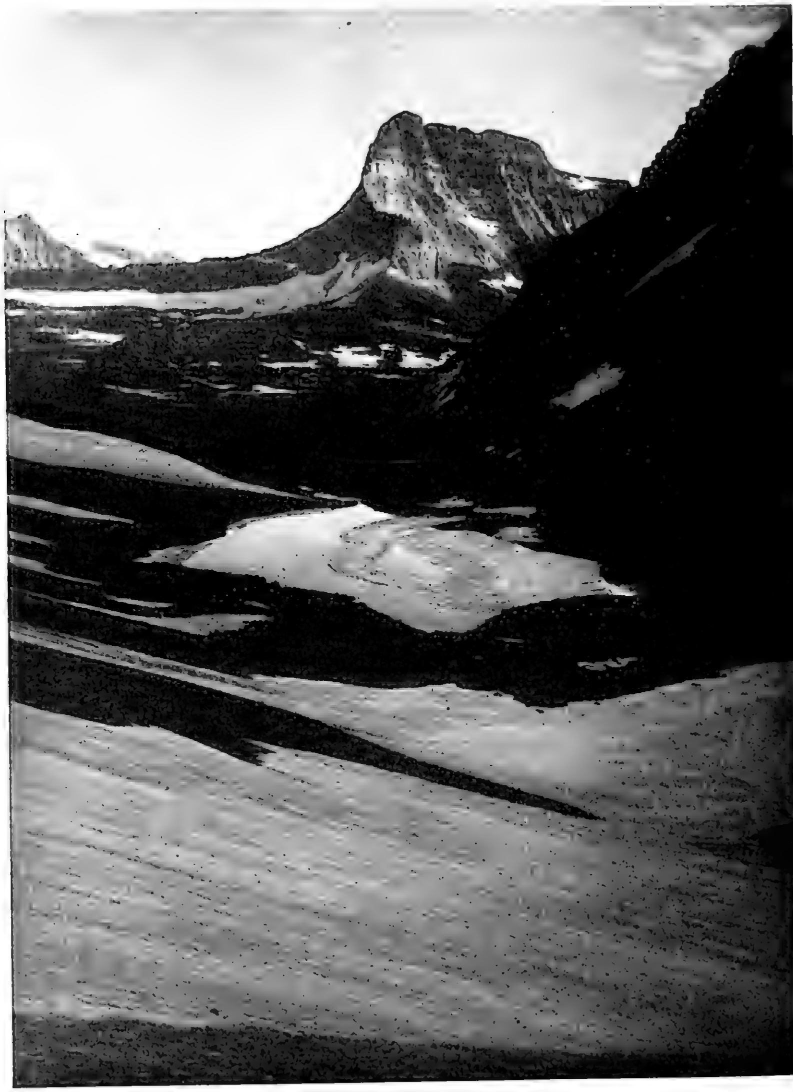

Fig. 8.—Monuments on either side of entrance to cirque above Ptarmigan Lake, Glacier National Park. (Photograph purchased of Northern Pacific Railway.)

The two districts which are here contrasted, the Bighorn Range and the Glacier National Park, furnish also the opportunity to contrast the effects of rock structure in modifying the forms of relief shaped by mountain glaciation. Whereas the high upland of the Bighorn Range has a core of massive rock, thus resembling the Wasatch and Uinta ranges and the Alps, the rocks of Glacier National Park are sediments and dominantly shales and limestones. It was to be expected that the characteristic structures of these

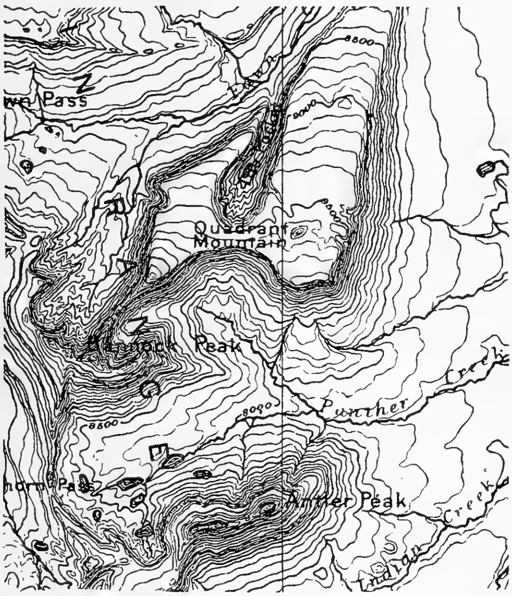

FIG. 9.—Cirque at head of Panther Creek, Yellowstone National Park, with pair of monumented peaks at entrance—Antler Peak and Bannock Peak.

sediments, their bedding planes and their joint system, should exert a strong influence upon the topographic forms produced, as indeed they have. The influence of the bedding planes is displayed in the Glacier National Park in the accentuation of the rock terraces at the upper ends of valleys within the cirques themselves. As in the Canadian Rockies across the international boundary, this character reaches an extreme (Fig. 11). An excellent instance is shown also in an illustration by Alden.[1]

[1] *Glaciers of Glacier National Park*, Fig. 11.

To the well-developed jointing found in the rocks of the Glacier National Park must be ascribed the well-marked checkerboard pattern displayed by the park valleys, a pattern which strikes one at once when the topographic map is examined. A number of observations of the bearings of master-joints which were made by

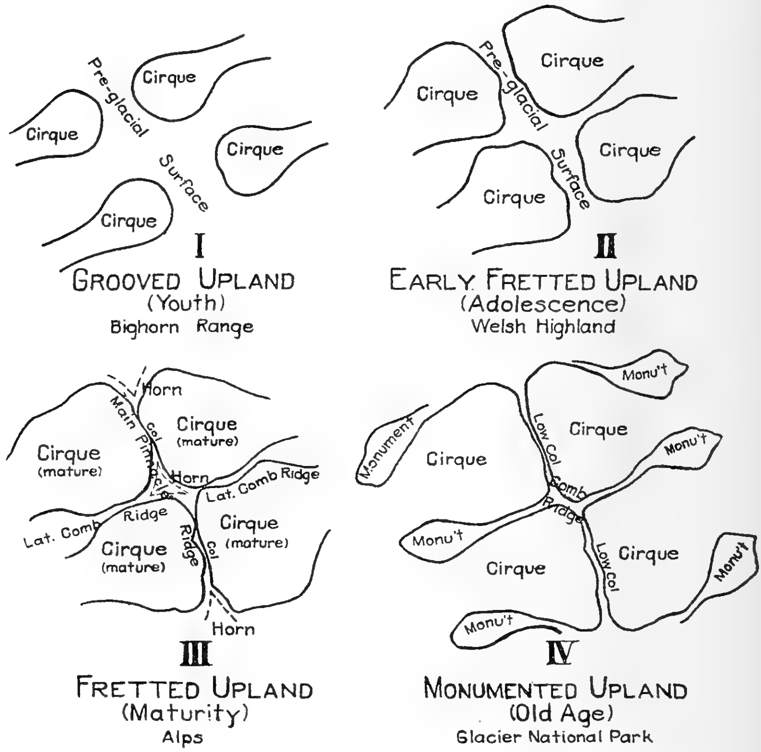

FIG. 10.—Stages of sculpture by mountain glaciers

the writer indicated a rather general correspondence between them and the trends of the valleys in which they were found. In some instances the lower spurs which have been less extensively sapped by glacial erosion indicate very clearly the dominating influence of the joint planes in shaping them. The cirques themselves also display this tendency by their approach to rectangular outlines.

II. THE TRANSITIONS BETWEEN THE MOUNTAIN GLACIER AND THE ICE CAP

From the standpoint of the sculpturing of the lithosphere, the ice cap is sharply set off from all types of mountain glacier through its inability to accomplish a sapping of rock surfaces due to rapid frost-weathering. Its sculpturing processes are therefore restricted to plucking, abrasion, and to a very limited extent frost-weathering on flattish surfaces—processes which in combination leave the rock rounded and presenting surfaces which are flatly convex

FIG. 11.—View of terraced cirque above Lake Grinnell, Glacier National Park

skyward. That these processes combined play but a subordinate rôle to frost-weathering in the case of all the types of mountain glaciers, would seem to be sufficiently attested by the sharply accented features which are brought about with their concavities toward the sky.[1]

Since the mountain glacier owes its very existence to a rock container within the lithosphere surface, the inclosing rock walls

[1] Hobbs, *Earth Features*, etc. (1911), p. 379, Fig. 405.

must in general project above the ice of the glacier. The rock surface will also be reached by air and water wherever crevasses descend through the ice of the glacier to the bed upon which it rests. The conditions essential to the sapping process are a supply of water on the rock surface and oscillations of temperature about the freezing-point. These conditions are not realized either in the case of ice caps or of continental glaciers, save only where nunataks emerge from beneath the ice near to the glacial margin.

When during an advancing hemicycle of glaciation a mountain glacier is so amply nourished that the rock walls of its containing basins become entirely submerged (ice-cap stage), a profound and immediate transformation takes place in the sculpturing processes.

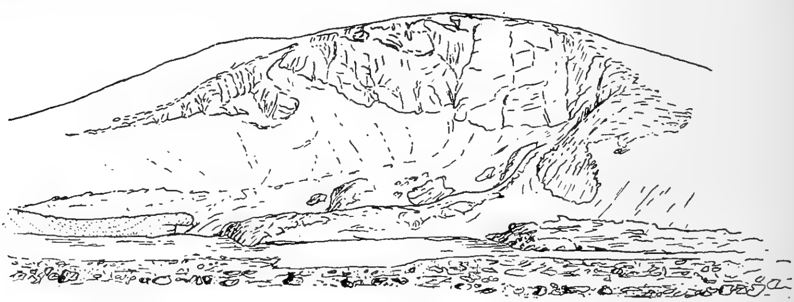

FIG. 12.—The northern cirque (Kjedel) on Galdhöpig in the glaciated surface of Norway. (After E. Richter.)

Up to this time, under the dominating influence of the sapping process, the effect of the glacial sculpture has been to sharpen all projecting features of the relief as the glacial basins and channels are carved deeper and extended outward from each individual locus. Now, however, under the plucking and grinding processes alone, which have usurped the functions of the frost-weathering, the pinnacles and horns within the comb ridges are truncated and ground down, with the result that above the shallowed cirques and the largely obliterated U-valleys there extends a flatly convex surface like that which is fashioned by the same processes beneath a continental glacier. The sharp relief which was inherited from the period of mountain glaciation is thus gradually ironed out into a flatly convex surface which is everywhere ground and polished by abrasion. The U-valleys are first effaced, beginning at their

lower extremities, and the last of the hollowed features of the inherited surface to disappear are the increasingly truncated remnants of the cirques, which in their later stages take the form of an armchair-like depression. Such features are well displayed in Norway where the continental glacier has similarly ironed out the inherited grooved or fretted upland (Fig. 12).

Such a surface as succeeds to a fretted upland under the sculpturing action of either an ice cap or a continental glacier will resemble in form a grooved glacial upland of extreme youth such as is illustrated by the Bighorn or Uinta ranges of the Rocky Mountain region, but it has less pronounced relief and, unlike such a preglacial remnant ("biscuit-cut" surface), *the upwardly convex surfaces are here planed and polished by abrasion.*

In the receding hemicycle of glaciation which succeeds to the culmination of glacial alimentation, the flat dome of the ice which constitutes the ice cap will have its surface progressively lowered until the stage is reached at which the rims of the buried cirque remnants begin to emerge from beneath their mantle of ice. In West Antarctica, near the winter quarters of the Swedish Antarctic Expedition of 1901–3, ice caps now blanket both James Ross and Snow Hill islands, and, like all Antarctic glaciers, they are in a receding hemicycle of glaciation. On the first-named island the rims of the cirques have emerged from beneath their cover along the eastern and southern margins of the island (Fig. 13). The Gourdon and Rabot glaciers are already apparently in large part detached from the dome of the ice cap, which here rises to its highest point in the Haddington berg. In the largest of the cirques lies the Hobbs Glacier, which is still in part fed by two ice cascades situated near the middle of the rim.[1]

Except that the continental glacier, and not an ice cap, has been the modeling agent, Mount Washington in the White Mountains[2]

[1] Otto Nordenskjöld, "Die schwedische Südpolar-Expedition und ihre geographische Tätigkeit," *Schwedische Südpolar-Expedition, 1901–3*, Vol. I, Lieferung 1 (1911), pp. 154–55, Map 3 and Pl. 13, Fig. 1.

[2] J. W. Goldthwait, "Following the Trail of the Ice Sheet and Valley Glacier on the Presidential Range," *Appalachia*, Vol. XIII (1912), pp. 1–23 (reprint), Pls. 1–9; "Glacial Cirques Near Mt. Washington," *Am. Jour. Sci.*, Vol. XXXV (1913), pp. 1–19; "Remnants of the Old Graded Upland on the Presidential Range of the White Mountains," *ibid.*, Vol. XXXVII (1914), pp. 451–53; "Glaciation in the White Mountains of New Hampshire," *Bull. Geol. Soc. Am.*, Vol. XXVII (1916), pp. 263–94, Pl. 13.

and Mount Ktaadn in Maine[1] would appear to supply near parallels to the sculpture just described, since the "gulfs" of the districts have been clearly recognized as cirques. Tarr has claimed that the mountain glaciers which sculptured the cirques on Mount Ktaadn

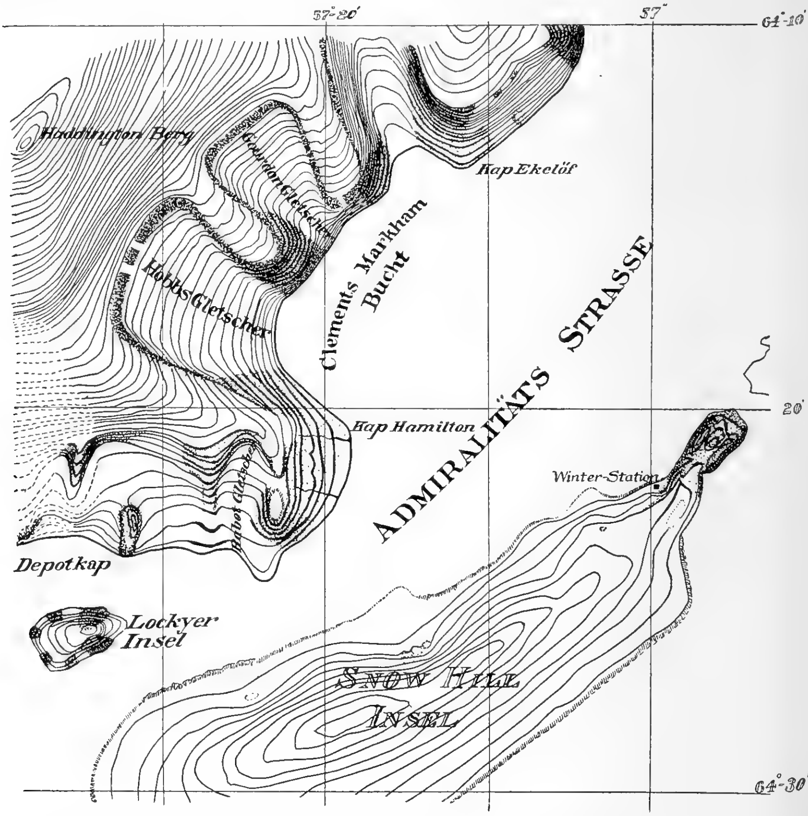

FIG. 13.—Map of portions of the James Ross and Snow Hill Islands of West Antarctica. (After Otto Nordenskjöld.)

were subsequent to the continental glaciation of the region. This is disputed by Goldthwait, who brings forward evidence to prove that in the White Mountains the mountain glaciers were antecedent to the continental glaciation which shaped the higher and flatter rock surfaces. We hardly see how there could fail to be glacial

[1] R. S. Tarr, "The Glaciation of Mt. Ktaadn," *Bull. Geol. Soc. Am.*, Vol. XI (1900), pp. 433–48.

remnants in occupation of the cirques for at least a brief period while the continental glacier was withdrawing from the region. These would presumably develop in much the same manner as those already described on James Ross Island, but with differences which will be pointed out in the next section of this paper. Goldthwait is no doubt correct in believing that the mountain glaciers had a much longer life during the advancing hemicycle of glaciation and that the cirques were shaped at that time. It is even doubtful if any appreciable work of erosion or deposition was accomplished in the later period of mountain glaciation, and this interpretation would be in harmony with Goldthwait's observations.

III. THE GLACIAL CYCLE ON THE MARGINS OF THE CONTINENTAL GLACIER OF ANTARCTICA

It is a fundamental and prerequisite condition for the sequence of stages through which mountain glaciers pass during a receding hemicycle of glaciation that the areas of alimentation and ablation should be sharply separated from each other. The former is restricted to the upper levels, and alimentation is augmented in amount toward the top, whereas the area of wastage is found in the lower levels and the losses are increased toward the bottom. Such a distribution results principally from two conditions: (1) mountain glaciers are nourished by upwardly directed air currents which deposit their moisture as a result of progressive adiabatic refrigeration; and (2) they are wasted by contact with warm-air layers whose temperature rises progressively toward the bottom. *It is a direct consequence of the combination of these conditions that mountain glaciers during a receding hemicycle of glaciation become reduced in area through withdrawal of the glacier foot up the valley, and even in its expiring stage the glacier head occupies essentially the same position that it did at the beginning* (Fig. 14).

Were these two conditions affecting the size of mountain glaciers not realized, the results would be quite different. When we examine the glaciers on the margins of the inland ice of the Antarctic, we find they differ widely from those of moderate latitudes, which are the ones that are well known and have formed the basis of our classification. Within the Antarctic air temperatures do not rise

above the freezing-point even in the summer season, save only during short intervals at the termination of the fierce Antarctic blizzards. Furthermore, these marginal glaciers to the inland ice are nourished, not by inwardly and upwardly directed air currents,

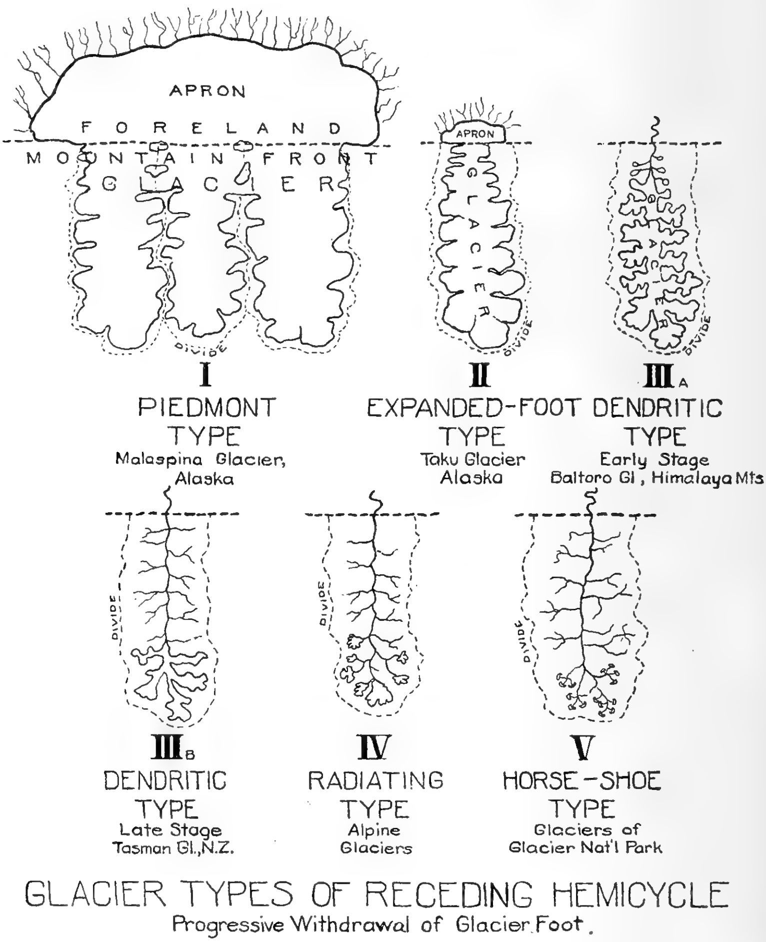

FIG. 14

as are the mountain glaciers of moderate latitudes, but by downwardly and outwardly flowing currents which bring drift snow from the inland ice and often carry it beyond the marginal glaciers to be dissipated upon the surface of the sea. *Separate areas of*

FIG. 15.—Map of an area in South Victoria Land near the winter quarters of the last Scott Expedition, showing waning glaciers which are withdrawing at both their upper and their lower margins. (After Griffith-Taylor and others.)

nourishment and waste in distribution with reference to altitude are thus not realized, and the otherwise universal law of exclusive drawing in of the foot of the glacier during its waning stages does not hold.

That this is true is particularly well shown in the area of waning glaciers described as "ice-slabs" by Ferrar, the glacialist of the first Scott Expedition to the Antarctic, and fully mapped by Griffith-Taylor, Debenham, and Wright of the last Scott Expedition[1] (Fig. 15). On a far larger scale and related to a continental glacier rather than an ice cap, these dying glaciers represent a later stage than those marginal types which have already been referred to from West Antarctica—the Gourdon, Hobbs, and Rabot glaciers of James Ross Island.

By examination of the map (Fig. 15) it will be noted that these glaciers must in an earlier stage have been connected together as a piedmont which was then a part of the parent area of inland ice lying to the westward. From that continental glacier when detachment occurred the rims of the battery of remodeled cirques which rise west of the existing glaciers must have emerged from the ice mantle in forms not unlike those now seen on the margins of James Ross Island. *Their subsequent diminution in size has gone on through withdrawal both from the cirques and from the lower portions of their valleys—from both extremities toward a central position at a moderate altitude, where the last stand will be made before final extinction.*

The usual law of ablation regulated with respect to altitude here plays, therefore, no part, and it is evident that the reflection and consequent intensification of solar heat radiation in the neighborhood of exposed rock walls has here been the controlling factor in localizing the wasting process. This effect of exposed rock surfaces has been recognized for high latitudes by the observation of moats surrounding nunataks[2] and of the lateral streams beside glacier tongues[3]

[1] Robert F. Scott, *Scott's Last Expedition*, Vol. II, map opposite p. 198.

[2] "Characteristics of Existing Glaciers," pp. 169, 257, Pl. 33B.

[3] *Ibid.*, Pl. 25A.

Reviews

Two Gas Collections from Mauna Loa. By E. S. Shepherd. *Bull. Hawaiian Observatory*, Vol. XIII, No. 5, May, 1920.

This is a brief report by Dr. Shepherd of the Carnegie Geophysical Laboratory at Washington on two gas samples collected by Dr. T. A. Jaggar, Jr. The samples were taken near the edge of a flow of incandescent rough pohoehoe lava on the south slope of Mauna Loa. The gases were collected in vacuum tubes from a depth of 2 feet in a 2-inch crack in the lava surface. The lava at a depth of 3 feet was glowing and the estimated temperature at the point of collection was 300° C.

A condensation of water within the tube was noted immediately upon collection. The analyses showed that about 70 per cent by volume of the gas (computed at 1200° C. and 760 mm. pressure) was water, in which respect the gases closely resemble those of Kilauea. About 16 per cent was nitrogen and the remainder mainly SO_3, SO_2 and CO_2. The water cannot be explained as the result of oxidation of hydrogen by admixed air, as is shown by the nitrogen percentage. If all the nitrogen were assumed to come from admixed air, the oxygen in such a quantity of air would be insufficient to account for the observed water.

The evidence of these samples accords with the classic work of Day and Shepherd at Kilauea in demonstrating the abundant presence of water in certain volcanic gases.

The gases of Mauna Loa show a high degree of oxidation, i.e., they have been almost completely burned. In general, they show a high degree of similarity to the Kilauea gases although the latter are rather variable. Especially noteworthy at Mauna Loa is the abundance of SO_3—2 to 8 per cent.

E. S. Bastin

The Geology and Ore Deposits of the Virgilina District of Virginia and North Carolina. By Francis Baker Laney. Virginia Geological Survey, University of Virginia. (Prepared jointly by the Virginia Geological Survey and the North Carolina Geological and Economic Survey.) 1917. Pp. 176.

The Virgilina district which lies partly in Virginia and partly in North Carolina is one of the copper districts in the eastern United States

that has produced considerable tonnages of ore. The investigation covers an area of approximately 550 square miles, including parts of Charlotte, Halifax, and Mecklenburg counties, Virginia, and parts of Granville and Person counties, North Carolina.

The area is made up almost wholly of igneous and highly metamorphosed rocks. They include ancient metamorphic gneisses and schists, the origin of which is unknown; a sequence of volcanic rocks, both basic and acidic, and volcanic clastics of each type, together with much volcano-sedimentary material; intrusive rocks of both basic and acid types, such as gabbro, diorite, granite, and syenite; and different varieties of dike rock, especially diabase. There is a small area of red or brown sandstone of Triassic-Newark age. Except the intrusives, the sandstones, and the dikes the rocks are all highly schistose and gneissoid in texture.

This prominent schistosity of the rocks is probably the most obvious structural phenomenon of the district, although jointing is prominent and there is conclusive evidence of folding. There is little direct evidence of faulting, but the intense dynamic metamorphism of the district could hardly have occurred without causing a certain degree of faulting.

With the exception of a few mineralized areas in more or less epidotized zones of the true basic schist, where deposits of native copper or of cuprite occur, the ore deposits are found in well-defined fissure veins, which occupy fractures in the rocks—in some instances possibly fault planes. The rock in which the veins occur is basic in character—the Virgilina greenstone—having the mineralogical and chemical nature of andesite; but it is thought that the vein material, both ore and gangue, was derived from the granitic magma of the region.

The gangue minerals, exclusive of included fragments of schist, named in the approximate order of their abundance, are: quartz, calcite, epidote, chlorite, hematite, sericite, albite, and possibly other plagioclase feldspars in small amount, and pink orthoclase.

The ore minerals, named in the approximate order of their abundance are: bornite, chalcocite, native copper, malachite, azurite, cuprite, chalcopyrite, chrysocolla, klaprothite (?), pyrite, argentite, silver, and gold. Of these minerals, bornite (in part), chalcocite (in part), chalcopyrite (in part), pyrite, klaprothite, argentite, native copper, and gold are regarded as hypogene or primary; while a part of the chalcocite, bornite, and chalcopyrite, and all the native silver, cuprite, malachite, azurite, and chrysocolla are held to be supergene or secondary.

The author gives a description of individual mines and prospects. A good geologic map of the district is appended to the report.

R. A. J.

The Stratigraphy and Correlation of the Devonian of Western Tennessee. By CARL O. DUNBAR. State of Tennessee, State Geological Survey, Bull. No. 21, Nashville, Tenn., 1919.

This volume is a detailed statement of the stratigraphy and correlation of the Devonian rocks of the western valley of the Tennessee River. The long sequence of the Devonian strata exposed in this region, especially the presence of the Upper Oriskany, and the abundance of fossils, probably will make this the standard section of the Lower Devonian of the entire Mississippi Basin. The important paleontological aspects of the problem are well treated. Following is the sequence of the Devonian formations of western Tennessee, as given by the author:

Series	Group	Formation
Neo-devonian	Chautauquan	Chattanooga shale Hardin sandstone member
	Senecan	Break
Meso-devonian	Erian	
	Ulsterian	Pegram limestone Break Camden chert Break
Paleo-devonian	Oriskanian	Harriman chert Break Quall limestone
		Break
	Helderbergian or Linden	Decaturville chert Break Birdsong shale Break Olive Hill formation: Flat gap limestone Bear Branch limestone Pyburn Ross limestone Break Rockhouse shale

R. A. J.

The Geology and Coal Resources of the Coal-bearing Portion of Tazewell County, Virginia. By T. K. Harnsberger. Virginia Geological Survey, University of Virginia, Bull. No. 19. Prepared in co-operation with the U.S. Geological Survey. 1919. Pp. 195.

This report deals with the coal resources of Tazewell County in southwestern Virginia. The surface rocks in the coal district belong to the Devonian, Mississippian, and Pennsylvanian systems. All the commercially valuable coal is in the Pennsylvanian.

The most prominent structural feature of the area is the Dry Forks anticline. The Pocahontas syncline and other folds occur in the region. The coal area is bounded on the southeast by a series of thrust faults. The Tazewell County coal field originally extended to the southeast far beyond its present limits but folds and faults lifted the coal-bearing rocks of the region to the southeast far above those of the present field and they have been removed by erosion.

The total area of coal land is 696.5 square miles. The total thickness of the coal-bearing formations is about 2,800 feet, every portion of which is exposed in some part of the area. At least fifteen coal beds are 30 inches or more in thickness over territory of sufficient extent to justify mining. In general the coal is of good coking quality and has a high fuel value. Because of the extreme variability of the coal beds, plans for development should be preceded by careful geological examination. Complete descriptions of the various coal beds are given. Included in the report are both a topographic and a geologic map of the coal area.

R. A. J.

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Among articles to appear in early numbers of the *Journal of Geology* are the following:

The Physical Chemistry of the Crystallization and Magmatic Differentiation of Igneous Rocks. III and IV. By J. H. L. Vogt.

Cycles of Erosion in the Piedmont Province of Pennsylvania. By F. Bascom.

Theoretical Considerations on the Genesis of Ore-Deposits. By R. H. Rastall.

Note on a Possible Factor in Changes of Geological Climate. By Harlow Shapley.

The Horizontal Movement of Geanticlines and the Fractures near Their Surface. By H. A. Brouwer.

Field Observations in Northern Norway Bearing on Magmatic Differentiation. By Steiner Foslie.

Geologic Reconnaissance in Baja California. By N. H. Darton.

Pleistocene Mollusca from Northwestern and Central Illinois. By Frank Collins Baker.

The Character of the Stratification of the Sediments in the Recent Delta of Fraser River, British Columbia, Canada. By W. A. Johnston.

Outline of Pleistocene History of the Mississippi Valley. By Frank Leverett.

Pennsylvanian Stratigraphy of North Central Texas. By Raymond C. Moore and Frederick B. Plummer.

Suggestions as to the Description and Naming of Sedimentary Rocks. By A. J. Tieje.

The Marine Tertiary of the West Coast of the United States: Its Sequence, Paleogeography, and the Problems of Correlation. By Bruce L. Clark.

The Relation of the Physical Properties of Natural Glasses to Their Chemical Composition. By William O. George.

Notes on a Recent Collection of Fossil Plants from the Dalles Group of Oregon. By Ralph W. Chaney.

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THE

JOURNAL OF GEOLOGY

JULY-AUGUST 1921

DIASTROPHISM AND THE FORMATIVE PROCESSES

XIV. GROUNDWORK FOR THE STUDY OF MEGADIASTROPHISM

PART I. SUMMARY STATEMENT OF THE GROUNDWORK ALREADY LAID[1]

THOMAS C. CHAMBERLIN
Research Associate, Carnegie Institution of Washington

PART II. THE INTIMATIONS OF SHELL DEFORMATION

ROLLIN T. CHAMBERLIN
The University of Chicago

PART I. SUMMARY STATEMENT OF THE GROUNDWORK ALREADY LAID

INTRODUCTION

As set forth in the first of this series of articles, it has been their main purpose to develop into more explicit form the basal ideas that logically belong to an earth built up of planetesimals. Inevitably, the alternative ideas that have been based on the older concept of an earth of gaseo-molten origin have been more or less constantly compared with them. The whole of the field has not yet been covered, but as the study now passes to a new and difficult phase, it is felt that it will be serviceable to assemble in brief, rather categorical statements such of the basal ideas already

[1] Published by permission of the Carnegie Institution of Washington.

developed as will form the groundwork for this next stage of the work, which will be a study of the megadiastrophism of the earth.

The introduction of the new term, megadiastrophism, calls for an explanation, if not an apology. It has already been found desirable by dynamic geologists to introduce the word "diastrophism," as a term of more comprehensive meaning than "deformation." The latter, while general enough in its etymological sense, has come to have a rather special meaning, by reason of its long usage to designate folding, faulting, and similar *declared* distortions of strata. It is not usually understood to denote those more intimate changes of form that take place in the deeper interior of the earth's body. To try now to make it include these would be at the risk of misinterpretation. But the new term diastrophism may be used to cover *any form of distortion of solid bodies*, and thus meet an imperative need.

There now arises a need for a still more comprehensive term which shall denote the diastrophism of the earth as a whole—or of large parts of it, such as continents and suboceanic segments—in a *collective* way without regard to the various special modes by which the diastrophism is effected. In the very nature of the case, the diastrophism of these great units will be composite and very complex, but we need to deal with them in a unitary way despite this complexity. The term megadiastrophism seems suitable for this purpose.

One of the most formidable obstacles in the way of bringing into actual use a new set of concepts where an old set has long had full possession of the thought, lies in the difficulty of really clearing the mind of *all the incidental factors* of the old concept, and of putting in their place *a full set* of the new. The difficulty does not lie so much with the main bold features of the new view as with the less obvious progeny of *derivative* concepts that must, in consistency, go out with the parent concept. In the study of any basal subject that has run far back into one's past thinking, a large brood of derivative concepts is quite sure to have been drawn out, but their connection with the parent idea is quite likely to have become obscure or to have passed entirely out of consciousness, so that the setting aside of the parent idea does not automatically

take them with it. It is not at all surprising, therefore, that, even in the most sincere endeavor to give true shape to a new issue under a new view for the purpose of a candid and hospitable test, some of the derivatives of the old view should unconsciously slip in and be treated as though they were offspring of the new. This, in reality, vitiates the whole test. The problem that has thus actually been fashioned and put to trial is a hybrid; it is not a true problem under either the old or the new view. For example, under the theory of a gaseo-molten earth, it was logically assumed that each spherical layer of the earth's interior was homogeneous; and hence it had a definite "melting-point."[1] There followed closely the inference that the material of such a layer must have a common state, either liquid or solid. From these logical derivatives of the primary assumption, far-reaching inferences were drawn in perfect consistency, and these, by years of association, have been woven into the web and woof of current thought with little consciousness that they are only dependencies of a cosmological postulate.

But if, on the other hand, the material of all such layers is very heterogeneous chemically, because it is an intimate mixture of planetesimal débris laid down at random, the logical inference is that each layer embraces *a wide-ranging group of solution temperatures* and has no single point of liquefaction. If it is subjected to a rising temperature, this would, at any given stage, cause the liquefaction of *only that fraction* of the material which was susceptible of liquefaction at the temperature reached, *not the "melting" of the whole layer*. This fraction would naturally be scattered throughout the mass of the layer and would give rise only to interstitial liquidity. The solution temperatures of the larger portion of the layer would not yet be reached, and this portion would remain solid. Now if the working mechanism of the body is so actuated by the joint force of internal and external stresses that graded pressures are brought to bear, greater below than above, and more or less intermittent, the disseminated liquid is likely to be kneaded out of the layer in the direction of least resistance and so leave the residue solid.

[1] In revised terms, as applied to interior conditions, a solution temperature or a narrow group of temperatures at which the constituents enter into mutual solution.

Now it will be seen that this second chain of derivatives is very different from the preceding chain and that the two are mutually exclusive. The links of the two cannot be mixed without the loss of all logical force. If mixed, the terms of the problem become a hybrid of incompatibles; such a chain does not exist in nature; it is not a real problem at all; it is merely a supposititious combination of incongruities.

The concept of isostasy gives rise to one set of derivatives, if based on the hypothesis of a crust floating on a liquid substratum inclosing a centrosphere of concentric homogeneous layers; and to quite a different set of derivatives on the hypothesis of a solid elastic earth whose internal material is heterogeneous and has suffered internal distortion.

The problem of the saltness of the sea has one set of subconcepts if the hydrosphere, at the outset, was as great as now or even greater, and quite another set if the hydrosphere started from a minimum and has grown steadily ever since and is growing still. To mix these throws the whole effort out of court. And so of not a few other earth problems of the more complex order.

In view of the difficulties of meeting the imperative requirements of consistency in working out such complex problems as the inner diastrophism of the earth, it is hoped that the following effort to reduce to brief convenient form the essential concepts already reached in the study of planetesimal accretion will be found serviceable. They are not a formal summary of the preceding articles nor drawn exclusively from them; some of them even have no dependence on the planetesimal hypothesis; they are merely found to be tributary to a satisfactory concept of megadiastrophism under the conditions of accretion. To make the statements brief and convenient, qualifications have been largely neglected and some statements may seem somewhat too baldly affirmative, but recurrence to the fuller discussions will, it is hoped, show that reasonable recognition has been made of legitimate grounds of doubt and needs of qualification. It is quite impossible here to accredit these propositions to those who have done most to develop them; they are merely assembled as propositions tentatively accepted as groundwork for further study.

GENERAL PROPERTIES OF THE EARTH

1. The solid elastic nature of the earth is accepted as having been put beyond serious question by the concurrent testimony of seismic waves, the body tides, the polar nutations, and collateral evidences.

2. The outer and major part of the earth is held to be minutely heterogeneous in chemical and physical composition, but yet, in a mechanical sense, sufficiently homogeneous to transmit seismic vibrations in legible form.

3. Some questions remain respecting the earth core; it is held to be dense and rigid; but earthquake waves traversing it do not, as yet, tell an unequivocal story. Possibly it is formed of concentric zones rather homogeneous in themselves but varying from one another; possibly, also, segregation of metallic matter toward the center has gone far enough to give a higher ratio of density to elasticity in this inner part than in the accretional zone above, and thus introduced seismic anomalies.

4. So far as now deducible, elasticity and rigidity increase toward the center faster than density. Since simple density segregation would scarcely carry a relative *increase* of rigidity and elasticity, this seems to imply a dynamic cause. The mean rigidity of the earth seems to be distinctly higher than that of steel.[1]

5. The major pulsation-period of the earth seems to be of the order of an hour but it is not yet precisely deducible from elastico-rigid-density data;[2] nor is the naturalistic evidence conclusive.[3] It may be near enough to commensurability with the semi-diurnal tide to strengthen it by resonance, but this is not certain.

6. The mode of motion of the poles in the earth is an index of high elastic rigidity. Schweydar has recently determined the

[1] Schweydar's recent determination is two and one-half times the rigidity of steel; he gives the rigidity of the central part of the earth as ten times that of the surficial part. "On the Elasticity of the Earth," *Naturwissenschaften* (1917), Potsdam, Germany, Part 38.

[2] On the influence of gravity on elastic waves, and in particular on the vibrations of an elastic globe, see T. J. A. Bromwich, *Proc. Lond. Math. Soc.*, Vol. XXX (1899).

[3] Nagaoka tried to deduce it from the eruptions of Krakatoa, *Nature* (May 26, 1907), pp. 89–91.

nutation period as 432.8 days.[1] This is so nearly commensurate with the fortnightly tide that there may be a resonance relation between them. The subnutation that has an annual period is probably the result of the seasonal shift of solar effects north and south of the equator.

7. The elastic nature of the body tide is accepted as practically demonstrated by the researches of Michelson, Gale, and Moulton, added to those of previous investigators.[2]

8. The water tides are held to spring in part from the body tide and in part from the direct attraction of the tide-producing bodies; their rise to notable value depends on the resonance of their basins.

THE TIME FACTOR

9. The arguments once urged against any great age of the earth because of the sun's short life, tidal action, etc., are held to be wholly invalid. An age somewhere between one billion and several billions of years seems best to fit in with astronomical and biological considerations. Ample time should be allowed for the evolution of star-clusters and the stellar galaxy, as well as life-evolution.

10. Estimates of the earth's age, based on current geological processes, require large corrections for the accelerating effects of present high reliefs and soil cultivation; in particular, for (*a*) increased vertical circulation, (*b*) more rapid cycles of evaporation and precipitation, (*c*) greater instability of vegetal clothing, (*d*) more rapid run-off, (*e*) deeper penetration of solvent action, (*f*) greatly increased soil waste, and (*g*) the much greater length of the low-relief periods than of the high-relief periods.[3] The required corrections are probably great enough to reconcile the geologic with the radioactive estimates.

11. For the computations used in the articles here summarized, a range wide enough to cover the uncorrected geologic as well as the radioactive estimates was used, as follows: (*a*) for the time since

[1] W. Schweydar, *Naturwissenschaften* (1917), Potsdam, Germany, Part 38.

[2] A. A. Michelson and Henry G. Gale, "The Rigidity of the Earth," *Jour. Geol.*, Vol. XXVII (1919), pp. 585–601.

[3] "The Quantitative Element in Circum-Continental Growth," Article VIII, *Jour. Geol.*, Vol. XXII (1914), pp. 516–28.

the beginning of the Paleozoic, from one hundred million to four hundred million years; (*b*) for that since the beginning of the Proterozoic, from three hundred million to twelve hundred million years; (*c*) for that since the earliest Archeozoic whose age has been estimated, from four hundred million to sixteen hundred million years.[1]

12. The time occupied in the evolution of terrestrial life is regarded as one of the most dependable evidences of the earth's age, though its testimony is of a rather general nature. Since the evolution from the early Paleozoic to the present is confessedly only a small part of the whole evolution, it was taken as 1/10 in the computations.[2]

13. Combining biologic, geologic, and radioactive estimates, the total period of life-evolution is taken roundly as lying between one billion and four billion years.[3]

14. The length of the period during which the rate of planetesimal infall was compatible with life, *previous* to the earliest determined Archean, is thus made to range between six hundred million and twenty-four hundred million years.[4]

15. Making allowance for the formative stage that preceded life-evolution, the whole age of the earth is taken tentatively as falling somewhere between three billion and five billion years.

All these estimates are of course only intended to serve working purposes in the light of the latest evidences; the whole matter is to be kept *sub judice* awaiting further light.

CONSIDERATIONS ADVERSE TO HIGH ESTIMATES OF DIASTROPHISM

16. Great thicknesses of shallow-water sediments do not necessarily imply great sinking of the crust. Measured in the usual way, thicknesses much greater than any observed may be laid down in the normal process of continental outgrowth without necessarily involving any crustal sinking at all.[5] Very thick

[1] "The Bearings of the Size and Rate of Infall of Planetesimals on the Molten or Solid State of the Earth," Article XIII, *Jour. Geol.*, Vol. XXVIII (1920), pp. 675–77.

[2] *Ibid.*, p. 675. [3] *Ibid.*, p. 676. [4] *Ibid.*

[5] "Foreset Beds and Slope Deposits," Article VI, *Jour. Geol.*, Vol. XXII (1914), pp. 271–72.

sediments, however, usually carry evidences of actual sinking, but such sinking must be determined on its own specific grounds.

17. Great sea-transgressions of the land do not necessarily involve great continental depressions; they are more or less due to general denudation, to shore cutting, and to sea-rise caused by sediment from the land.[1]

18. Effective base-leveling implies an absence of crustal movement while it is in progress.[2] It thus bears on the promptness and completeness of isostatic readjustments.

PERIODICITY OF DIASTROPHISM

19. Effective base-leveling is evidence that the earth-body is strong enough to stand the strain of ordinary loading and unloading for a long period without essential yielding; it thus implies that isostatic adjustments are periodic rather than continuous, and that diastrophism, in so far as it is assignable to such loading and unloading, is similarly periodic.

20. The stages occupied in base-leveling and sea-transgression were probably much longer than the intervening stages of active deformation.

21. The mechanism of isostasy implies that great basins once formed tend to remain basins permanently, and that great protuberances tend to remain protuberances except as worn down. Isostasy is not hospitable to great inversions of sea and land. To this there may be regional exceptions where great erosion is closely paralleled by great deposition.

22. Since the present isostatic status follows a period of great diastrophic readjustment, it is an open question whether the present degree of compensation is essentially a consequence of that diastrophism, or is a normal state approximately maintained at all times.

[1] "The Lateral Stresses within the Continental Protuberances and Their Relations to Continental Creep and Sea-Transgression," Article III, *Jour. Geol.*, Vol. XXI (1913), p. 585; "Rejuvenation of the Continents," Article IV, *Jour. Geol.*, Vol. XXI (1913), pp. 673–75.

[2] "The Rejuvenation of the Continents," Article IV, *Jour. Geol.*, Vol. XXI (1913), pp. 676–81.

23. The data assembled by Rollin Chamberlin[1] imply that prolonged loading and unloading are either (*a*) sufficient to cause diastrophism at long intervals, or else (*b*) act as triggers to set off other forces that were steadily accumulating during the quiescent periods. Whether loading and unloading are in themselves sufficient causes of deformation or not is a question on which studies in megadiastrophism are expected to shed decisive light.

THE TOTAL AMOUNT OF SELF-COMPRESSION OF THE EARTH

24. In spite of the foregoing conservative considerations, the amount of unequivocal deformation shown in Paleozoic and later strata alone is so large as to overtax all resources of diastrophism safely assignable to the cooling of the earth, and yet the very complex diastrophism of the earlier areas greatly exceeded this later diastrophism. The deformations since the middle of the Miocene are so great—in view of their lateness in the history of the earth—as to suggest that the causes of diastrophism are very persistent and very profound.

25. If we try to measure the total diastrophism by comparison with the total life-evolution, a result more than ten times that of the Paleozoic and later ages is implied, for diastrophism should have been very active in the formative ages and declined afterward, while life-evolution appears to have been accelerated as time went on.

26. If the intimate crumpling, close folding, and faulting of the Archean and Proterozoic terranes is made the basis of estimate, the total diastrophism must have been very great, but just how great it is impossible now to say.

27. If the present continents be looked upon as the outcome of a contest between sea-shelf outbuilding, on the one side, and inthrust from the oceans, on the other, the total diastrophism is clearly large, but very difficult to estimate definitely.

28. If the early earth be supposed to have been segmented in a natural mechanical way, and the existing continents and ocean basins interpreted as derivatives from these segments by

[1] "Periodicity of Paleozoic Orogenic Movements," Article VII, *Jour. Geol.*, Vol. XXII (1914), pp. 315–45.

outgrowth, thrust, and deformative shift, the total diastrophism appears even greater than that inferred from the preceding data.

The concurrent import of all lines of evidence is that the total diastrophism of the earth was very great, but a more comprehensive and quantitative mode of estimate is needed, and especially one that covers the diastrophism of the formative stages, in respect to which these lines are very weak.

DIASTROPHISM ESTIMATED BY PLANETARY COMPARISON

29. A comparative study of the volumes and densities of the earth and its neighbors affords an entirely independent mode of estimate and gives definite quantitative results.[1]

30. If the moon were built up of moon-stuff—having its present mean density of 3.34—to a sphere whose mass equaled that of the earth, it would have a volume of 430,353,000,000 cubic miles, while the actual earth's volume is only 259,924,000,000 cubic miles. To reduce the hypothetical moon-earth to the volume and density of the real earth would require a shortening of the radius of 725 miles and of the circumference of 4,555 miles.[2]

31. If Mars were built up of Mars-stuff at its present mean density of 3.58 to a spherical body of the mass of the earth, it would have a volume of 401,502,000,000 cubic miles. To compress this to a body of the density and volume of the earth would involve a shortening of its radius of 618 miles and of its circumference of 3,883 miles.

32. If Venus were similarly built up of its own material to a mass equal to that of the earth, its volume of 289,506,000,000 cubic miles would have to be shrunk radially 177 miles, and circumferentially 1,112 miles, to have the volume and density of the earth.

33. The earth, moon, Mars, and Venus revolve in a tract whose total width is less than 3 per cent of the radius of the planetary system. They were, therefore, probably formed under much the same dynamics, of much the same kinds of material, and in much

[1] "The Order of Magnitude of the Shrinkage of the Earth Deduced from Mars, Venus, and the Moon," Article X, *Jour. Geol.*, Vol. XXVIII (1920), pp. 1–17.

[2] *Ibid.*, p. 13.

the same way, and hence are closely comparable. There might have been some gradation of material, but since the earth is compared with the next outer and the next inner planet, any such gradation is largely equated in combining the results. These four bodies may therefore be taken as representing four stages of growth of a single body under the conditions that prevail at their mean distance from the sun, which is substantially the earth-distance.

34. The giant gaseous planets cannot properly be compared with these solid bodies without radical qualification, for the giants were probably gaseous from the start and never underwent the sifting necessary to cull out material unsuited to form the earth and kindred bodies (see 48 below). The evolution of the giant planets belongs to a distinctly different category.

35. Comparing the densities of the earth, Venus, Mars, and the moon as they now are, a marked increase of density with increase of mass is shown: to wit, the moon, with mass 0.0122 (earth = 1), has a density 3.34 (water = 1); Mars, with a mass 0.1065, has a density 3.58; Venus, with mass 0.807 (?), has a density 4.85 (?); and the earth, whose mass is unity, has a mean density 5.53.[1]

36. Closer inspection shows not only an increase of density with mass but *an accelerated rate of increase of density for each increment of mass.* This clearly implies that *their densities arise* from their own massiveness, an inference in harmony with No. 4 above.[2]

37. Under the kinetic theory of gases, the larger the mass, the greater its ability to hold light molecules. Greater proportions of intrinsically light matter were therefore almost certainly gathered into the more massive bodies. *They became dense in spite of a larger proportion of intrinsically light material.*[3]

38. Let it be noted that the acquirement of high density is not held to be a matter of simple mechanical compression; this was a conditioning factor in the process, but only that. There were probably added (*a*) progressive rearrangements and reorganizations of the material into denser forms as the stresses grew, including not only simple physical readjustments but the formation

[1] *Ibid.*, p. 10. [2] *Ibid.*, pp. 16–17. [3] *Ibid.*, p. 17.

of new compounds, new minerals, and possibly even new molecules and new atoms; (*b*) an increase of endothermic compounds as the temperature increased; (*c*) the removal of liquefied material and its heat of liquefaction; as also (*d*) the self-heating radioactive substances that helped to produce the liquidity. This process was essentially a metamorphic one, but in the deep interior the stresses and temperature rose to a much higher order than in the zone of observation, and the metamorphism is assumed to have been more radical.[1]

39. Simple pressure experiments are incompetent to determine the limit of self-compression in this broader sense. Such experiments cannot even cover the full range of metamorphic reorganization recorded in the observational zone. They are much less competent to set metes and bounds to the higher order of metamorphism under immensely greater stress conditions. At best they can only indicate how much of the observed effect is to be assigned to simple mechanical compression and how much to metamorphism in the interest of higher density.

40. To set the new view into sharp contradistinction to the old, let it be noted that the planetesimal earth is looked upon as a profoundly metamorphic earth with a minor igneous accessory, while the traditional earth is commonly regarded as a profoundly igneous earth with a minor surficial metamorphic subsidiary.

RESCRUTINY OF THE FORMATIVE PROCESSES

The importance of the results of planetary comparison made it seem imperative to rescrutinize the whole chain of postulates that lies back of them. Chief among these were the tenets of the planetesimal hypothesis itself. This hypothesis was therefore reconsidered from the ground up. At the same time competing hypotheses were retested, for these lie, in a similar way, back of the older views of diastrophism, whether their authors are aware of it or not. This restudy brought forth not only evidence confirmatory of previous views but supplementary considerations which need to be summarized here as part of the groundwork on which further study is to proceed.

[1] *The Origin of the Earth* (1916), chap. ix, "The Inner Organization of the Earth," pp. 226–40.

41. To follow accurately the logic of the planetesimal hypothesis, it is necessary to keep clearly in mind that its point of view is pre-eminently *dynamic*. Its type ideas are not based on *material forms*, but on *dynamic organizations*. This was set forth in explicit terms in the earliest full statement of the hypothesis.[1] In spite of this, the hypothesis has come to be more or less unconsciously regarded as dependent on a special interpretation of spiral nebulae, because these nebulae have been much used as illustrations of the type of deployment supposed to have been involved in the genesis of our planetary system, but the theory is not thus dependent, as was urged from the outset. It will stand or fall solely on its ability to explain the remarkable characteristics and relationships of our planetary system. The requirements imposed by these are so many and so exacting that no theory but the true one has any chance of fully meeting them. We may, of course, think they are met when they really are not. We hold it is quite sure, however, that in time the "vestiges of creation" will give convincing tests, and the essentials of the whole history of the earth will be read from beginning to end.

42. Two distinct types of planetesimal organization are recognized: In one, planetary nuclei, serving as collecting centers, revolve among the planetesimals and gather them in, forming bodies of notable size. In the other, there are no such collecting nuclei, and the formation of bodies of planetary size is a practical impossibility; the planetesimals remain small and constitute a multitude of minute secondaries.

43. Two distinct classes of planetesimal secondaries are recognized: (*a*) those which the parent body may develop *by its own genetic resources*, i.e., *monoecious* secondaries, and (*b*) those which can be developed only by the *co-operation* of another body serving as a second dynamic parent, i.e., *dioecious* secondaries.[2]

[1] In outlining the planetesimal hypothesis for the use of students in Chamberlin and Salisbury's *Geology*, Vol. II (1905), pp. 38–40, it was specifically pointed out that the planetesimal condition may arise in different ways, as from a gaseous nebula of the Laplacian type, or a meteoritic swarm of the Lockyer type. An origin from a spiral nebula was made the leading type for reasons specified, but to this was added: "While this will be followed as the type view, let it be distinctly noted that the planetesimal doctrine of accretion does not stand or fall with this particular conception" (p. 40).

[2] *The Origin of the Earth* (1916), "Celestial Kinships," pp. 101–2.

Monoecious secondaries may arise as orbital ultra-atmospheres[1] do, and in similar ways, and are normally minute. They thus probably attend all stars in prodigious numbers but small mass. Dioecious secondaries are assigned to the dynamic action of a passing body on an eruptive sun by first stimulating an effective outburst and then drawing the projected matter into orbits about the mother-body, a purely dynamic function. The eruptions are quite sure to take place by successive impulses, and a part of each belch is likely to remain under self-control and act as a collecting center for the more scattered planetesimal part. Thus the main mass will be gathered into a comparatively few, rather large planets.

44. While thus *monoecious* systems may be nearly universal, *dioecious* systems can arise only when the necessary dynamic encounter takes place. Close approaches of stars are rare events; there is, therefore, no reason to suppose that planetary systems *like our own* are common in the heavens. Only one such is known. Still, rarely as one star closely approaches another, the multitude of stars and the great length of celestial time makes possible a fairly large number of even this very peculiar class of secondaries. It is a logical error to base arguments on the assumption that planetary systems of this type are universal or even necessarily frequent attendants of stars.

45. The cosmological rescrutiny brought out in stronger terms than before the extreme improbability that a planetary system like our own would arise from any form of centrifugal action in a condensing gaseous or quasi-gaseous nebula. The principles of the kinetic theory of gases, combined with the dynamic considerations that lie back of the Roche limit[2] and of the new criterion of Moulton,[3] indicate that all such action would result in minute secondaries without effective collecting nuclei, a condition which practically inhibits the formation of large planets.[4]

[1] *The Origin of the Earth* (1916), "Celestial Kinships," p. 21.

[2] F. R. Moulton, "On the Application of Roche's Limit and a New Criterion of Somewhat Similar Character," *Astrophys. Jour.*, Vol. XI (1900), pp. 120–26.

[3] *Ibid.*

[4] "Selective Segregation of Material in the Formation of the Earth and Its Neighbors," Article XI, *Jour. Geol.*, Vol. XXVIII (1920), pp. 137–44.

46. Planetary generation by the dioecious method is not confined to the approach of one *star* to another; bodies less massive than stars, if they make sufficiently close approaches, are competent to develop planetary systems. Only 1/745 part of the mass of the sun was required to form the planets of our system. The critical point in such cases is the ability of the small passing body to impart the requisite revolutionary momentum.[1]

47. New potentialities of projection from the sun have recently been disclosed. Twice during 1919, Pettit observed that erupted calcium vapor ascended by *a succession of accelerating impulses.* In one case, the ejected calcium vapor increased its outward velocity from 5.5 kms. per second to 60 kms. per sceond; in the other, from 37 kms. per second to 163.9 kms. per second. In both cases, the calcium was moving at its highest velocity when it ceased to be visible, high above the sun, probably either from cooling or from scattering, or both.[2]

DIVERGENCIES IN THE MODES OF PLANETARY CONDENSATION

48. The planetary nuclei diverged into two lines of descent almost as soon as they emerged from the sun. The nuclei that were massive enough to remain hot and gaseous at all stages, and to hold practically all molecules that came within their control, naturally grew to be giant planets. Nuclei that were not massive enough to hold all the solar gases, but in the main only the heavier ones, such as later made up the stony and metallic bodies, followed a much more selective career. This was a very vital matter, for the solar gases, constituted as they were, could not condense directly into bodies of the composition of the earth; a preliminary sifting was indispensable.[3]

49. A further divergence soon followed in this sifted class. A few of the larger nuclei were only incompletely sifted; so that they retained relatively small amounts of gases of the atmospheric

[1] "Multiple Phases of the Planetesimal Hypothesis," *ibid.*, pp. 149–50.

[2] Recent Disclosures Bearing on the Solar Parentage of the Planets," *ibid.*, pp. 145–49; Edison Pettit, "The Great Eruptive Prominences of May 29 and July 15, 1919," *Astrophys. Jour.*, Vol. L (October, 1919), pp. 206–19.

[3] "The Physical Phases of the Planetary Nuclei during Their Formative Stages," Article XII, *Jour. Geol.*, Vol. XXVIII (1920), pp. 481, 487–89.

type. In most cases, however, the nuclei were unable to hold any appreciable amount of such gases. Some of the smallest could not hold the hot vapors of even stony and metallic substances. This would have led to their complete dissolution but for the fact that, on emergence from the sun into interplanetary space, they promptly condensed into *clouds of precipitates and these soon gathered into precipitate aggregates.* These, being much larger and less active than the molecules of the previous vapors, were held under control and collected into planetoids.[1]

50. The formation of precipitates and precipitate aggregates of stony and metallic substances apparently played an important part in the condensation of planetary nuclei. As such precipitates appear to be forming now in the photosphere of the sun, it is assumed that they would be formed freely in solar gases projected into planetary space. Such aggregates would act as Brownian particles and the condensation would not be strictly gaseous. If the nuclei later passed into the liquid state, crystalline and concretionary aggregates would probably form and give rise to a solid-liquid Brownian mixture. The descent was therefore that of Brownian mixtures of different types rather than simple gaseous condensation.[2]

51. Each planetary nucleus must have inherited internal motions from its solar state and from its ejection, and this must have promoted cooling and precipitation during the first critical stages. Later, convectional movements were added and continued the precipitation. The inherited motions must have been more or less asymmetrical and this tended to give asymmetry to the core as it solidified.[3]

52. The inherited motions and the sifting processes were sources of hazard to each small nucleus. Probably the smallest planetoids and satellites now seen were the smallest that could be formed in this way. The spheres of control of even small nuclei were, however, surprisingly large, and this was doubtless the saving

[1] "The Physical Phases of the Planetary Nuclei during Their Formative Stages," Article XII, *Jour. Geol.*, Vol. XXVIII (1920), pp. 492–98.

[2] "The Formation of Precipitates and of Brownian Mixtures," *ibid.*, pp. 489–92.

[3] "The Motions Inherited from the Solar Eruption," *ibid.*, pp. 483–87.

factor. The sphere of control—as against the sun at the distance of the earth—of a mass 1/20 of that of the earth is 458,000 miles in diameter. A mass no greater than 0.000,000,296 of the mass of the earth has a sphere of control 8,200 miles in diameter (MacMillan). The functions of spheres of control in the genesis of planets are a feature that has been too much overlooked.[1]

53. Asymmetry in a planetary core was likely to make itself felt in the distortion of the mass later built upon it. The inherited notion that the earth core, if liquid, would be merely a melt whose solidification would have to await the progress of freezing from its surface downward belongs to an old order of thought. The solidifying process should rather be studied as progressive supersaturation and precipitation.[2] The formation of the solid core doubtless depended chiefly on the order in which the various possible minerals were formed and the extent to which they settled toward the center. The inherited and convectional motions doubtless affected the lodgment of the precipitates and rendered the growth of the core more or less asymmetrical.

54. Such external forces as gave rise to changes in the rate of rotation, or produced tides or nutations, must have played their part in distorting the forming core.[3]

THE SIZE OF THE PLANETESIMALS

55. At the outset, the planetesimals were merely the molecules of the scattered solar gases or the minute precipitates formed from these or else the molecules which escaped later from the nuclei. Starting thus small, and conditioned by their wide dispersion, their growth was necessarily not only slow but precarious.[4]

56. The Zodiacal Light is reflected by minute particles that probably have orbits of the planetary type, and hence are in fact

[1] "Table of Dynamic Properties," *ibid.*, p. 478. For function of spheres of control, see Article XI, *ibid.*, pp. 128–34.

[2] "The Critical Conditions That Controlled the Passage of the Nuclei into Collecting Cores," Article XII, *ibid.*, pp. 477–500.

[3] "Exterior Agencies That Affected the Planetary Cores during Their Formation and Afterward," *ibid.*, pp. 500–504.

[4] "The Nature of the Planetesimals at the Start," Article XIII, *ibid.*, pp. 666–71.

planetesimals. If they are planetesimals left over from the formation of the planets, they are surely old enough to have grown to the largest practicable sizes; but, in spite of this, they are certainly quite small. If they have had a more recent origin, they should still represent normal growth. In any case, they add their testimony to the smallness of planetesimals.[1]

57. The restudy of cosmological processes led to a new view as to the relations of the erratic elements of the solar system, the meteors, meteorites, and comets, to the normal elements, the planets, planetoids, planetesimals, satellites, and satellitesimals, to the effect that they were all formed by the same type of dynamic action, save that the former were given erratic orbits, while the latter were given concurrent orbits. The ways in which this difference arose are given in the original discussion. These give a unitary view to the whole solar system. Now under them, chondrules are interpreted as aggregates of stony and metallic precipitates from solar gases, growing in a manner similar to that of the planetesimals. If so, the sizes of chondrules and planetesimals should be about the same. Chondrules vary in size from walnuts down to dust particles, millet seed being mentioned as representative. Their history differs in some features from that of meteorites, and they might be called meteoresimals to distinguish them. They seem to be immensely more numerous than meteorites; probably at least a hundred million of these meteoresimals "burn out" in the upper air as "shooting stars" for every meteorite that reaches the ground.

58. Meteorites proper are interpreted as fragments of erratic bodies disrupted by the extremes of heat and cold they suffer at the two ends of their very elliptical orbits. The parent bodies are held to have been formed by the aggregation of precipitates in a way similar to the aggregation of planetoids (No. 49 above), except that the original material was given very diverse and elliptical orbits.[2] Under this view, it is not the meteorites but the chondrules that are analogous to the planetesimals.[3]

[1] "The Zodiacal Planetesimals," Article XIII, *Jour. Geol.*, Vol. XXVIII (1920), p. 672.

[2] "The Testimony of the Aberrant Bodies of the Solar System," *ibid.*, pp. 696–701.

[3] *Ibid.*, pp. 697–98.

59. The size of planetesimals is not really important in considering the heating effects of their infall, for what might be gained by their aggregation into large sizes would be lost by the lessened frequency of their infall, but to round out the inquiry a study was made of the effects of the infall of supposedly large bodies.

60. The great pit and the scattered material of the famous Meteor Crater (Coon Butte), Arizona, the greatest of accessible examples of its kind, show that the impact of a large body—probably meteoritic or cometic—caused extremely little melting but yet great excavation, much upturning and wedging aside of strata and wide scattering of débris. Its lesson is that the impact of such bodies converts their energy of motion *mainly into new forms of mechanical work*, and very subordinately into melting.[1]

61. On the supposition—not in fact accepted—that the craters of the moon are pits formed by infalling bodies, the proper inference from the observable effects is the same as that from Meteor Crater. The steep-walled pits and the great radiating lines of débris are direct evidence of great mechanical effects. The lofty walls of the pits show no signs of the collapse that should attend great melting. The evidence of lava flows, in proportion to the number and size of the craters, is not remarkable on the most favorable interpretation. The level tracts, interpreted as lava, may be merely débris plains. The lunar Alps, Apennines, and other mountain ranges of the moon imply at least a crust stout enough to sustain such great elevations. Nothing observable definitely implies a holo-molten state.[2]

MELTING EFFECTS OF PLANETESIMAL INFALL

62. The planetary nuclei undoubtedly picked up some planetesimals that had wide-ranging orbits, but the larger portion specially related to the earth were distributed in the form of a ring around the sun about 55,000,000 miles in breadth, 58,000,000 miles in depth, and 292,000,000 miles in length. All these planetesimals had their own independent orbits, and were sustained by their own revolutionary energy. Their condition was antithetical to the collisional and collapsing habit of molecules in a gaseous

[1] "The Testimony of Coon Butte or Meteor Crater," *ibid.*, pp. 686–89.

[2] "The Questionable Intimations of the Craters of the Moon," *ibid.*, pp. 690–94.

organization. They were subject to collision with one another, indeed, but their sparse distribution made the contingency less immanent than might easily be imagined. When collisions did occur, rebounds were more probable than mutual coherence; permanent unions were only probable when their relative speeds and other conditions were specially favorable. After union, they were liable to be driven apart again by succeeding collisions. Ultimate capture by the earth core was probable, but even in this case only as their orbits favored. They were perturbed by all the attractions of the solar system. These, on the whole, favored capture by the earth nucleus but not in all cases. The process of collection was intricate, indirect, and slow.[1]

63. As a step toward realizing the sparseness of the planetesimals and the time required for their collection, let them be supposed to stand still as at first distributed, while the earth nucleus, taken as a net 6,000 miles in diameter, sweeps through them at 18 miles per second. To expedite the work, let the path of this net be so shaped and shifted by some demon that it will clean up an entirely new swath at each revolution. Even then it would take about 100,000,000 years to sweep up all the planetesimals.[2]

64. To try, as a next step, the most rapid natural way, let the planetesimals act as though particles of a gas, collapsing on the track of the nucleus after each sweep—though that is far from their habit—and let each sweep, as before, clear a path 6,000 miles in diameter at the rate of 18 miles per second. It would then take about 260,000,000 years to gather in 90 per cent of the planetesimals; to sweep up all would require an indefinite period.[3]

65. The real case was much less favorable. The planetesimals and the nuclei were moving in the same general direction, at somewhat similar speeds. They could thus come together, as a rule, only as one overtook the other, or as their paths converged, a relatively slow method. They attracted one another but this did not necessarily bring them together, for their orbits might become so adjusted to one another that they merely became traveling companions, like the earth and the moon. Mutual attractions

[1] "The Intimations of the Planetesimal Mechanism," Article XIII, *Jour. Geol.*, Vol. XXVIII (1920), pp. 677–78.

[2] *Ibid.*, p. 678.

[3] *Ibid.*, p. 679.

would constantly give rise to perturbations and these, on the whole, would favor the gathering in of the planetesimals, but just how fast is beyond the reach of rigorous computation. It can only be reached roughly by approximations. A period somewhere between one billion and three billion years seems most probable.[1]

66. While the total heat of planetesimal infall was great, the melting effects were conditioned by the rate of fall and by the extent of atmospheric surface into which they fell. At the stage when the earth was one-third grown, there would remain to fall in 4×10^{26} planetesimals averaging 1/50 lb.; or about 3×10^{15} planetesimals per square foot of earth surface. Taking the radio-bio-geologic estimate (2.4×10^{9} years), as the time, the average rate of plunge into the upper part of each one-square-foot air-column would be one planetesimal in 6.7 days. Even at the time estimate based on the older geologic scale—probably much too short—a planetesimal would fall into the top of each square-foot air-column not oftener than once in 40 minutes. Now in the upper air, half the heat would be promptly radiated outward, and the rest would act at a disadvantage in heating the earth's surface some miles below. The mean intervals between falls would quite surely be much too great to produce the melting of the earth's surface.[2]

For a cross test, let the method be reversed by selecting a rate supposed to be sufficient to produce melting and testing this by the time. Let it be supposed that one fall per second per square-foot air-column would have been a melting rate. All the planetesimals would, at this rate, have been collected in a little over 4,000 years, an impossibly short time. It can scarcely be held that a rate of one fall per hour per column would produce surface melting; and yet, at that rate, all would have fallen in less than 15,000,000 years. We found that the impossibly speedy demon-directed method required 100,000,000 years. The melting of the surface during the last two-thirds of the earth's growth seems out of the question.[3]

[1] *Ibid.*, pp. 679–81.

[2] "The Rate of Planetesimal Infall," *ibid.*, pp. 681–86.

[3] *Ibid.*, pp. 683–86.

67. The protracted rescrutiny of the cosmological postulates back of the planetary comparison was made to see if the results of that comparison needed any serious modification or qualification. It was found that the total self-compression should be placed somewhat higher than the high figures given by the comparison, on account of the larger proportion of light material which the larger bodies gathered in. As the amount had previously been found to be unexpectedly large, the correction may be treated simply as a margin of safety. The trustworthiness of the comparative method seemed to be greatly strengthened by the rescrutiny; the deductions from the planetary comparison are therefore regarded as firm groundwork for further study.

THE DIASTROPHIC RESOURCES OF A PLANETESIMAL EARTH

68. An earth built of planetesimal dust settling from the air in a mixed state would retain, to an almost ideal degree, its latent resources for subsequent chemical combination and physical reorganization. It would retain also about as much as possible of its potential energy of position, for the accessions would be very loose as first laid down. In strong contrast to this, the resources of a molten earth would be dissipated in large measure while still in the fluid state. The molten globe spent its energies in a hot youth; the cooler planetesimal earth conserved its resources for its later life.

69. In a molten earth, the high heat would be the master factor; its rate of dissipation would set the pace of progress. In a planetesimal earth, the strength of the solid material would be the ruling factor. Self-compression would take place only as this was overcome. Heat, of course, would be developed, but only as the yield of the solid matter permitted; it would be merely an incident of the process and would help to stay the process until it was taken care of.

70. In a dust-built earth, self-compression began as soon as a new layer was laid on an old one. Thereafter, compression continued by stages and intervals as long as loading continued. Diastrophism therefore ran through the whole formative history and was doubtless more active during the stages of growth than since.

71. It seems inconsistent with the ultimate constitution of matter, as now understood, to assign any limit to compression so long as pressure increases and there is any way by which the energy of organization can escape or take a new form of greater density. As energy continues to escape from the earth, and as there still remain resources of reorganization into denser forms—at least in the outer part of the earth—diastrophism is probably yet far from the end of its career. It is probably competent to rejuvenate the continents for eras yet to come.

VULCANICITY AS A DIASTROPHIC AUXILIARY

72. Our planetary system, embracing nearly a thousand bodies all told, presents a great graded series in which the largest mass is many million times that of the smallest.[1] There is also a gradation in physical state. At the upper extreme, Jupiter is dominantly fluid; at the lower extreme, the planetoids and satellites are atmosphereless solids; in the middle, gases, liquids, and solids are combined. The earth is near the middle but dominantly solid. The dividing line, where fluids and solids might be supposed to be critically balanced, lies considerably above the earth in the series. The subordination of the fluid element in the earth is assignable to certain restraining factors imposed by the rigidity of the material. These give rise to a partition of the energy set free by self-compression, so that only a portion of it manifests itself as temperature. A portion becomes refixed in endothermic compounds; a portion is consumed as latent heat of liquefaction and is forced into higher horizons where a solid state is resumed and the liquefying energy is again set free and readily discharged, while a third portion is probably consumed in physical and, perhaps, even atomic reorganization. The joint effect is the persistent removal of liquidity and the conservation of solidity. The whole is a profound metamorphic process. The special processes of vulcanism are thus looked upon as subsidiary to the metamorphic-diastrophic processes, but still as important auxiliaries.

[1] "The Physical Phases of the Planetary Nuclei during Their Formative Stages," Article XII, *Jour. Geol.*, Vol. XXVIII (1920), table I, p. 476.

73. Since the formation of mutual solutions of rock material within the earth affects only such part of the mixed matter as becomes soluble under the contacts and conditions present, and since the solid state is resumed at or near the surface, *magmatic generation* is the matter of primary moment in the history as well as the philosophy of magmatization. Magmatic *differentiation* belongs essentially to the reverse process, and is dependent on the generative process.

74. The temperature curve of the interior, under this view, does not depend primarily on cooling from the surface or on the arrest of a convectional circulation, but on dynamic action within the body itself, starting with the restraints of inherited solidity in the clastic matter and adding new restraints at intervals later, by transformations of such a nature as to fit a part of the mixed material for a higher state of solidity, while a part was liquefied and sent to the surface to resume solidity there.

75. The gases and gas-producing substances entrapped by the burial of minutely mixed planetesimal matter should have been well-nigh a maximum. Subsequent processes of partial liquefaction and extrusion should have set these gases free, and they should have joined whatever liquid material was in process of formation. The magmas should thus have been rich in gases; sometimes becoming explosive. A large gaseous factor is therefore held to be characteristic of the vulcanism of an earth so built.

76. On the other hand, during the protracted boiling of a gaseo-molten earth, potential gases should have been set free to a maximum, and all gases should have been brought to the surface by convection, whence they should have escaped to the fullest extent consistent with gravity, because of their hot state. There should have remained in the boiled liquid merely the equilibrium quantity required to balance the partial pressures of the atmosphere.[1] Laboratory melts of like material under like conditions should indicate the limited amount of this. The cooled mass could scarcely have carried those abundant supplies of gas that have been so amply manifested by the extrusive action of all the

[1] Rollin T. Chamberlin, "The Gases in Rocks," *Jour. Geol.*, Vol. XVII (1909), pp. 565–68.

geologic ages. If the lunar craters are volcanic, as we assume, the evidence against a molten moon becomes still more imperative, for even in its cold, mature state the moon cannot hold free volcanic gases. All such gases should have escaped while the moon was still hot and boiling, and it should later have cooled to a smooth, gasless globe, singularly unfitted for the explosive action which its surface implies.[1]

CLUES FROM SURFICIAL DIASTROPHISM

77. The diastrophism of a solid earth should have been a unit, in all its great essentials. The deformations of the shell should have been intimately related to the diastrophism within the shell, if indeed not largely dependent on it. The mode of junction of the under surface of the shell with the upper surface of the interior mass should be especially instructive. One of the newer methods of study has disclosed the important fact that very notable downward protrusions are developed. These are defined by plunging zones of accommodation that are at least suggestive. The intimations of these are herewith added as Part II, since these seem to belong with this résumé of ground work for megadiastrophic study. Other studies in the zone of observation offer clues of great value; indeed, no line of inquiry lacks them. Two of these are very specially related to the study of inner diastrophism: the experiments of Adams[2] that point toward a higher degree of rigidity than was accepted previously, and the contributions of Van Hise[3] and Leith[4] to the methods of metamorphism, especially the selective and rejective phases of anamorphic action, which point toward methods of reorganization very like the more radical selective and rejective metamorphism assigned to the deep interior of the earth.

[1] Article XIII, *Jour. Geol.*, Vol. XXVIII (1920), pp. 694–95.

[2] Frank D. Adams, "An Experimental Contribution to the Depth of the Zone of Flow in the Earth's Crust," *Jour. Geol.*, Vol. XX (1912), pp. 97–118; F. D. Adams and J. A. Bancroft, "On the Amount of Internal Friction Developed in Rocks during Deformation, and on the Relative Plasticity of Different Types of Rocks," *Jour. Geol.*, Vol. XXV (1917), pp. 597–637.

[3] C. R. Van Hise, "A Treatise on Metamorphism," *Monogr. 47, U.S. Geol. Survey* (1904).

[4] Leith and Mead, *Metamorphic Geology* (1915).

PART II. THE INTIMATIONS OF SHELL DEFORMATION

As the result of field studies in Pennsylvania in 1905, the crustal shortening involved in the folding of the Appalachian Mountains west of Harrisburg was found to have been fifteen miles. Computations based upon this shortening and the consequent upbowing seemed to indicate that the shape of the deformed section was that of a triangular prism or wedge pointed downward.[1] Two sides of the wedge were found to converge beneath the mountainous tract till they came together under the middle portion of the deformed belt at a depth of thirty-two miles. No consideration whatever of stress-and-strain relations entered into the deduction of this wedge-shaped form. It came out directly from the graphic treatment of the field measurements, quite irrespective of any theory of mechanics, or of the nature of diastrophism; in fact, the result came as a distinct surprise. It was soon seen, however, that the shape was in harmony with the principles of fracture under essentially uniform horizontal compression.

The key to the method used in this inquiry lay in the axiomatic proposition that there must be a definite relation between the thickness of the deformed shell, the horizontal shortening which this shell has suffered, and the amount of resulting vertical bulge, on the assumption that there has been no notable compacting of materials.[2] When the horizontal shortening and the consequent upswelling have been determined from field measurements, the thickness of the deformed shell can readily be calculated. The most important outcome of the method is that it gives the under-configuration of the deformed shell in addition to other qualities. It is from this feature that the most important intimations relative to the deeper diastrophism are drawn.

Besides the under-configuration of the deformed shell, the following generalizations seem to be warranted: (*a*) Sharp folding

[1] Rollin T. Chamberlin, "The Appalachian Folds of Central Pennsylvania," *Jour. Geol.*, Vol. XVIII (1910), pp. 228–51.

[2] For a discussion of this and other possible qualifying factors, see *Jour. Geol.*, Vol. XVIII (1910), pp. 236–37, and also "The Building of the Colorado Rockies," *Jour. Geol.*, Vol. XXVII (1919), pp. 235–38, 244–47.

and much crustal shorting indicate the deformation of a thin shell; otherwise the resulting upward bulging would be enormous. (*b*) Open, gentle folding may signify either a thin shell or a thick shell, according as there has been little or much upbowing. (*c*) For a given crustal shortening, the greater the vertical uplift the thicker the shell which has been actively deformed.

The detection of the wedgelike shape of the deformed mass led to a consideration of the nature of the plunging planes which outline the block. If the wedge was defined by fracturing, the borders, as shown by the Daubrée experiment,[1] by the familiar crushing-strength tests upon building stones, as also by an analysis of the stress-strain relations, should be fault planes dipping beneath the deformed block at angles in the general vicinity of 45° though in most cases somewhat less. This is the result to be expected in a case of non-rotational strain, in which the axes of strain do not change position with respect to the axes of stress. If the developing strain be rotational in character, the angles of the shearing planes will be lowered from 45°, in proportion to the extent of the rotational element.[2] If, on the other hand, definite shearing planes do not develop, and the deformation is largely by folding, it is possible that the folding dies out below by affecting successively narrower and narrower belts. Though such a process would make the deformed block taper downward, just as in the preceding case, the borders would be much less sharply defined. With increasing resistance to deformation with increasing depth, in accordance with the results of the experimental work of Adams and his colleagues,[3] it seems mechanically logical that folds should die out in this fashion. But whatever the nature of the bordering zones of accommodation may be, the results of the computations in the case worked out show that the folded tract becomes narrower

[1] G. A. Daubrée, *Études synthétiques de géologie expérimentale*, T.I., p. 316, Plate II.

[2] R. T. Chamberlin and W. Z. Miller, "Low Angle Faulting," *Jour. Geol.*, Vol. XXVI (1918), pp. 1–44.

[3] Frank D. Adams, "An Experimental Contribution to the Depth of the Zone of Flow in the Earth's Crust," *Jour. Geol.*, Vol. XX (1912), pp. 97–118; F. D. Adams and J. A. Bancroft, "On the Amount of Internal Friction Developed in Rocks during Deformation, and on the Relative Plasticity of Different Types of Rocks," *Jour. Geol.*, Vol. XXV (1917), pp. 597–637.

below, thus maintaining the general wedge shape. The deepest folds lie beneath the middle of the belt.

In the broader study of diastrophism it is to be recognized that movement is not wholly confined to the strongly deformed masses, but takes place in some less degree in the masses that lie below them and at their sides. Associated movements are to be recognized which extend deep into the earth and possibly even throughout its whole mass. The relations between the distinctly distorted section and the environing portion are various. For example, the bordering thrust-faults on the margins of the southern Appalachians and of many other strongly compressed mountain ranges indicate that actual fracturing and shear take place very commonly between the strongly deformed and the slightly deformed blocks near the surface. Where the deformation has not been so intense, a sharp upturn of the strata in a great fold may mark the borders of the mountainous belt, as at Tyrone, Pennsylvania, and in the Colorado Rockies. Here actual fracture has not developed to any important extent, though there has been an approach toward fracture on the outer limbs of such folds. The adjustment between the more movable, more deformed portion and the less movable, less deformed region has been accomplished, partly by shearing and partly by mass rearrangements taking place in the folding process. With increasing depth below the surface, actual faulting should diminish, though distributive shear should presumably descend much deeper. No limiting depths can be assigned, for the time element plays an important part, though not easy to evaluate. To quick-acting stresses the earth reacts as an elastico-rigid body; under long-continued stresses it yields to slow mass movement. With greater depth molecular rearrangement and recrystallization should presumably take precedence. These might be manifested by folding, or by cleavage, or perhaps only by rock flow. Under such conditions the deformed block would probably not be sharply bordered.

While the border belts near the surface are in places actual fault planes, nevertheless, throughout most of their extent they probably constitute zones, perhaps of considerable breadth, separating the more deformed mountain mass from the less deformed

mass to the side of it and below it. The accommodation may take place by differences in the amount of wrinkling on the two sides, differences in the extent of elongation and schistosity, or differences in some less clearly defined type of rock flow. It may be merely a zone in which, as one goes from the undeformed region into the crumpled mountain belt, folding rapidly becomes more pronounced and an increase in schistosity becomes marked. The dividing belts may be vaguely outlined or they may be more sharply defined.

Because the wedge shape seemed to be in general harmony with certain recognized facts and principles, it was natural enough to suspect that it might prove to be a type of diastrophism of wide application. Wedge dynamics might prove to be characteristic of other mountain systems, and might be applied perhaps also to the elevation of plateaus, and those movements which control the rise of continental masses. A plateau-forming movement, if the outcome of lateral thrusting, would be assigned to a thick shell gently wrinkled. Very little shortening of a still deeper section would suffice to elevate a mass of continental dimensions.

In 1910, following the publication of the Appalachian paper, an attempt was made to apply these principles to continental diastrophism. Cross-sections of the globe were drawn with border planes dipping inward beneath the continents at 45°. Because of the curvature of the earth, each plane, in order to carry out the principle consistently, was drawn to cross the different radii of the globe at 45°. The result of such a treatment is shown in Figure 1. Outlined thus, the continents appear as shallow units very subordinate to the oceanic segments. The latter are truly the master segments, which squeeze the continental wedges periodically outward, as well as laterally, when the materials of the contracting globe become strained beyond their yielding point.

This suggested extension of the wedge principle to the continents was not pushed farther at the time, for the reason that certain possible objections quickly came to mind, so that it seemed advisable to allow the question to lie fallow for a while and await developments. After a wait of ten years, during which time this principle was discussed with several successive classes of graduate students,

and during which time geologic thought has progressed more and more toward the conception of a solid globe, it now seems worth while to put the ideas briefly into print. Objection will, of course, at once be made by many to any prolongation of shearing planes

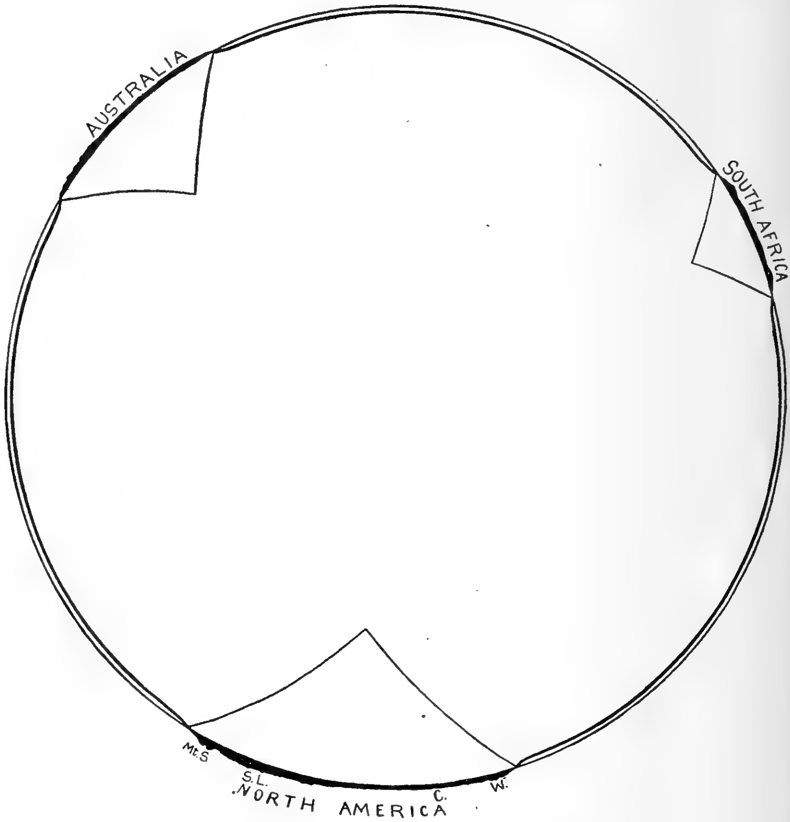

FIG. 1.—The continental wedges. Section through Washington, Chicago, Salt Lake, Mt. Shasta, Australia, and South Africa. Surface relief exaggerated and diagrammatic. Drawn in 1910.

deeply beneath the surface. The tenacity of the idea of easy flowage in the depths is remarkable. The idea of a molten interior, though disclaimed in name by most of those who follow closely the development of geologic science, is yet followed unconsciously, in fact, to some degree at least by most of them. On the other hand,

it is the view of those inquirers who accept the modern evidences of increasing solidity with increasing depth that the idea of easy movement in the interior is to be scrupulously avoided, whatever may be the form or mode of movement. It is held by them that differential motion of all kinds in the solid interior takes place only in response to high differential stress. If the term flow is to continue in use, it should carry simply the idea of an intimate distributive method of deformation and be shorn entirely of all suggestion of easy movement, for that belongs to liquidity. Strong support for this view is found in the experimental work of Adams, which extends the zone of cavities to greater depths than formerly supposed possible, owing to the increasing strength of the rocks under cubical compression.

There are increasing grounds for the view that the various special methods of deformation have a more intimate association with one another than has been generally recognized, and that processes heretofore confined to the upper zones may have application to greater depths. The true status of present knowledge of the movements in the deeper unseen zone has been most judiciously and trenchantly stated by Leith in his vice-presidential address before Section E of the American Association:[1]

> Notwithstanding these and other considerations, any conclusion as to the existence of a deep zone in which all rocks flow when deformed is hypothesis, not proved fact, and perhaps will always remain so. The environmental conditions are not accurately known; and even if each of the factors were measured, their conjoint effect is still speculative. Variations in the time factor alone may determine whether a rock flows or fractures. Rock flowage which has occurred in rocks now accessible to our observation fails to indicate increase with depth with sufficient clearness and definiteness to warrant confident downward projection.

The general purport of the under-configurations developed by these shell studies carries at least an intimation that the principles of surficial diastrophism are to some extent applicable to the deeper problems of the continents as well. When the ocean basins sink and the continents are uplifted, some adjustment necessarily takes

[1] C. K. Leith, "The Structural Failure of the Lithosphere," *Science*, Vol. LIII (1921), pp. 195–208.

place between the segments. Near the surface the adjustment may well be accomplished by thrust-faulting; by movement along distributed shear planes where clear-cut faulting does not develop; by accommodative folding, and in part by the deformation involved in the formation of flow cleavage, and possibly in other ways. It is important to understand clearly that the border zones between the more deformed blocks and the less deformed blocks are not held to be simple fault planes, though thrust-faults do so commonly emerge at the surface in these situations, but instead are held to be zones of more composite nature in which adjustments between the larger units are accomplished in the various ways outlined above.

In full agreement with the view that the results of deformation, in the zone of observation, afford perhaps the best criterion we have at present for judging of the behavior in the deeper zones, we may make the projecting planes, or zones, between the oceanic and continental segments the line of approach to the deeper problem, especially if the problems of the deeper horizons are as strictly the deformations of solids as are those of the surface, however different their conditions of temperature and pressure. With depth, fracture should become an exceptional phenomenon, though certain types of shear should be more persistent. And it would seem likely that deformation by distributive movement involving crumpling, cleavage, and general rock flow, should predominate in the deeper horizons.

Cleavage, by recrystalization, develops parallel to the elongation of the mass, whatever be the nature of the compressive stress which produces it.[1] From the experiments of Daubrée[2] upon the orientation of mica flakes under various conditions of compression, we learn that the direction of elongation is determined by the direction of least resistance. The deformation is controlled by the difference between the stresses along greatest and least axes of stress, and in this the least axis of stress is most important. By changing the position of the orifice (direction of easiest relief) in a compression cylinder, such as that used by Daubrée, *without changing*

[1] C. K. Leith, *Structural Geology* (1913), pp. 84–87.

[2] *Op. cit.*, pp. 407–22.

in the least the direction of applied force, the mica flakes could be made to assume any desired orientation. This illustration is introduced here for the reason that geologists so commonly refer to the direction of applied force as though it of itself determined the result, and relate everything to this direction without considering with equal care the lines of resistance. But it is the differential stress which is all important in deformation, and in this the axis of least compressive stress plays a part of the most critical importance. In fact, "lines of least resistance" might well be made the topic of a hortative sermon.

When a region is subjected to strong compressive stress under ordinary conditions, the axis of least stress is the vertical one, and the easiest relief is upward. Elongation then presumably takes a vertical direction, as does also whatever flow cleavage develops. But under conditions of special burden, lateral elongation may be a condition precedent to a final vertical one. A notable shear zone extending obliquely downward on the 45° principle, such as is here postulated between segments, might be expected to exert an orienting effect on the direction of most ready relief for some distance beyond the point where actual shear ceases, though it is uncertain how far beyond. After that influence ceased to be effective, the inclination of the ensuing schistosity should theoretically become more nearly vertical. If, in the deeper parts, the border planes between the segments come into parallelism with the elongation under recrystallization, and so with the schistosity, they should become steeper below. But it is not certain that the parallelism exists, nor do we know the controlling conditions sufficiently well to be certain that the elongation of the mass will be straight upward.

In the actual drawing of the border planes, the angle of 45° serves largely as a convenient average inclination, suggested by the planes of no distortion in the ellipsoid of strain.[1] It is, however, recognized in engineering practice that the angle of fracture under compressive stress, even where the strain is entirely non-rotational, varies widely from 45°, depending upon the nature of the

[1] C. K. Leith, *Structural Geology* (1913), pp. 16–20.

substance.[1] This variation applies especially to rupture, and it is uncertain how deep rupture may be safely projected. In rock flowage other conditions obtain. The inclination might be expected to become steeper. If it be true that in the lower reaches the border zones should become more steeply inclined than here drawn, that would amount to projecting the roots of the continental masses to greater depths. But the real significance of the location and behavior of the bordering zones in the deeper portions of the globe can only be satisfactorily treated by tracing the diastrophic phenomena throughout the stages of the earth's growth. Under the plantesimal view, each of the stages involved at first surficial diastrophism and later underwent the various deeper diastrophisms involved in the upgrowth of the continental and oceanic segments. This will be taken up in a later paper of this series.

According to the general philosophy of which the wedge theory is a part, condensation of material in favor of greater density takes place throughout the deep interior under the influence of gravitative force. This causes shrinkage and the development of strong lateral thrusting in the outer portion of the globe. When the growing stresses reach and exceed the strength of materials under the conditions obtaining within the earth, a period of diastrophism sets in. The vaster and heavier oceanic segments take the lead in descending and as they do so, the continents, several or all, are wedged upward. Some may be wedged up more than others, or one side of a continent uplifted more than the other sides. A moderately thick shell forming only part of a continent may suffer notable shortening and be pushed up into a plateau. A thinner shell along the borders of a continent, yielding more readily and suffering much greater shortening, may be folded and faulted into a mountain system. The tangential compressive stresses necessarily extend throughout the outer portion of the globe and are not to be thought of merely as thrusts from an active oceanic mass against a passive continental mass, but the actual deformation into mountain systems is, for reasons to be brought out later, a

[1] Walter H. Bucher, "The Mechanical Interpretation of Joints," *Jour. Geol.*, Vol. XXVIII (1920), pp. 707–30; Vol. XXIX (1921), pp. 1–28.

distinctly localized phenomenon occurring where the compressive stresses exceed the strength of materials.

In conclusion, it is felt that the shell-deformation is intimately related to the less obtrusive diastrophism of the subshell mass. The plunging zones that form the common border of the interlocking tracts give suggestive intimations that may have wider application and more profound significance. These form a line of approach to the larger diastrophic problem and so properly constitute a part of the groundwork for the study of megadiastrophism.

THE PHYSICAL CHEMISTRY OF THE CRYSTALLIZATION AND MAGMATIC DIFFERENTIATION OF IGNEOUS ROCKS

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II

$CaMgSi_2O_6$ (DIOPSIDE) AND SiO_2

The binary eutectic between $CaMgSi_2O_6$ (melting-point = 1391°) and SiO_2 (melting-point = *ca.* 1700°, see p. 330) consists, according to Bowen (*Amer. Jour. Sci.*, Vol. XXXVIII [1914]) of 84 per cent Diops:16 per cent SiO_2, with melting-point = 1362°, and at this temperature SiO_2 forms tridymite.

THE TERNARY SYSTEM $CaMgSi_2O_6$:Ab:An AND THE SEQUENCE OF CRYSTALLIZATION BETWEEN THE PYROXENE MINERALS AND PLAGIOCLASE

The ternary system between the chemically pure substances $CaMgSi_2O_6$, $NaAlSi_3O_8$, and $CaAl_2Si_2O_8$ has recently been explained by N. L. Bowen in a treatise "The Crystallization of Haplobasaltic, Haplodioritic, and Related Magmas" (*Amer. Jour. Sci.*, Vol. XL [1915]).

The melting-points are: $CaMgSi_2O_6$, diopside = 1391.5°. $CaAl_2Si_2O_8$, An = 1550 ± 2°. $NaAlSi_3O_8$, Ab = 1100 ± 10°.

The binary eutectic Diops:An was determined as 58 per cent Diops:42 per cent An, with melting-point at 1270°.

On account of the extremely high viscosity of the melting masses, consisting of predominant Ab, the binary eutectic between Diops and Ab could not be determined, but by extrapolation (see Fig. 6) it lies at only a few per cent of Diops to nearly 100 per cent Ab, and naturally at a temperature lower than the melting-point of Ab, consequently somewhat lower than 1100°, thus a couple of hundred degrees lower than the eutectic between Diops and An. From the theoretical explanation of Schreinemaker

it appears (cf. p. 337) that the ternary system Diops:Ab:An must form two melting-planes, one for Diops and the other for Ab+An. This is verified by Bowen's experimental investigations, whereby further is proved that the eutectic line between

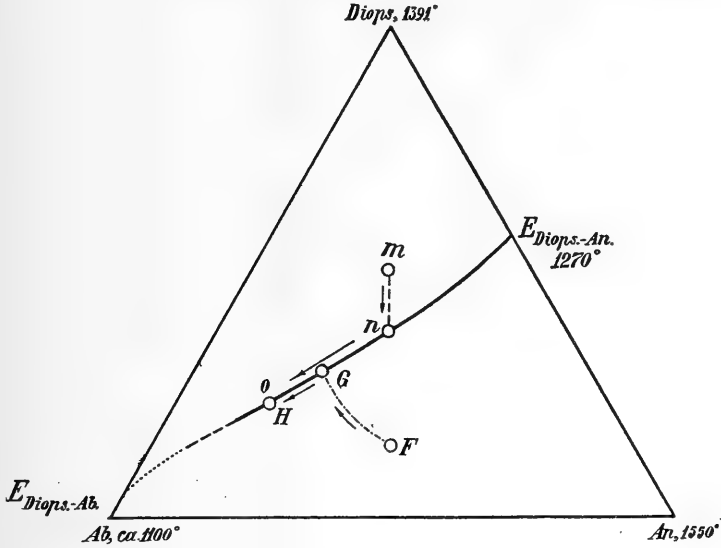

FIG. 6.—The ternary system An:Ab:Diops (horizontal projection), after Bowen

$E_{Diops.-An.}$ and $E_{Diops.-Ab.}$ has a *continuous* decline, without a maximum or a minimum. From Bowen's diagram (Fig. 6) I calculated some points on the eutectic line.

			Melting-Points
58 Diops:42 An: 0 Ab	or	58 Diops:42 $An_{100}Ab_0$	1270°
51 Diops:39 An:10 Ab	or	51 Diops:49 $An_{80}Ab_{20}$	*ca.* 1260°
44 Diops:36 An:20 Ab	or	44 Diops:56 $An_{65}Ab_{35}$	*ca.* 1238°
38 Diops:32 An:30 Ab	or	38 Diops:62 $An_{50}Ab_{50}$	*ca.* 1235°
33 Diops:27 An:40 Ab	or	33 Diops:67 $An_{40}Ab_{60}$	*ca.* 1225°
28 Diops:22 An:50 Ab	or	28 Diops:72 $An_{30}Ab_{70}$	*ca.* 1215°
23 Diops:17 An:60 Ab	or	23 Diops:77 $An_{22}Ab_{78}$	*ca.* 1205°
18 Diops:12 An:70 Ab	or	18 Diops:82 $An_{15}Ab_{85}$	*ca.* 1185°

The sequence of crystallization is illustrated by some examples taken from Bowen's treatise, to which, however, I add a few comments. In a melt (*m*) of 50 per cent $CaMgSi_2O_6$ and 50 per cent $Ab_{50}An_{50}$—accordingly with a surplus of $CaMgSi_2O_6$—diopside

commences to crystallize, if subcooling is left out of consideration, at a temperature of 1275°. After the separation of a certain amount of diopside the melting mass arrives at a point (*n*, *ca.* 38 per cent Diops:62 per cent $Ab_{50}An_{50}$ and at a temperature of 1235°) on the eutectic boundary curve. Then a simultaneous crystallization of diopside and plagioclase commences, the latter in the beginning with a composition $Ab_{20}An_{80}$, but on continued cooling with continually less An. On the presumption of a *complete equilibrium* between the solid and liquid phases, accordingly between the already crystallized plagioclase and Ab+An in the solution, the composition of the crystallized plagioclase is continuously displaced. The quantity of the liquid grows continually less by continuous crystallization. The last remnant of liquid is spent at a point (*O* on Fig. 6) 23 Diops:60 Ab:17 An, at the melting-point 1200° and with a separation of a minimal quantity of plagioclase $Ab_{50}An_{50}$.

If, on the other hand, we choose a melt of 15 per cent $CaMgSi_2O_6$ and 85 per cent $Ab_{50}An_{50}$—consequently with a surplus of Ab+An—plagioclase of the composition $Ab_{20}An_{80}$ commences crystallizing at the temperature 1375° (see point *F* on Fig. 6). On continuous cooling at first only plagioclase crystallizes. On the presumption of a *complete equilibrium* the composition of the already separated plagioclase is continuously changed. In this manner a plagioclase Ab_1An_2 appears at 1300°. The eutectic boundary curve is reached at *G*, at a temperature of 1216°. Now a simultaneous crystallization of diopside and plagioclase commences, as the Ab:An relation in the separated plagioclase little by little is displaced. At 1200° the last remnant of liquid is spent, the entire mass of plagioclase having the composition $Ab_{50}An_{50}$.

With *lacking equilibrium* the first separated plagioclase remains unchanged. By the continuous separation of relatively An-rich plagioclase, the quantity of Ab in the solution increases continually, and even more strongly than by the equilibrium. In this manner, by the simultaneous crystallization of diopside and plagioclase along the eutectic boundary-line we here at last obtain a plagioclase with a very high percentage of Ab.

Bowen emphasizes—and rightly—that in a ternary system Ab:An:Diops (as in the analogous system Ab:An:Qu) there is

no eutectic *point*. But an eutectic *boundary curve* exists with *simultaneous* crystallization of plagioclase and diopside. This simultaneous crystallization does, however, not take place at a constant temperature, but continues some distance at decreasing temperatures. With complete equilibrium this drop of temperature is, however, relatively small, as by the examples above chosen: for 50 per cent Diops: 50 per cent Ab_1An_1, from 1235 to 1200°, and for 15 per cent Diops: 85 per cent Ab_1An_1 from 1216 to 1200°.

In deep-seated igneous rocks (with very slow solidification) a complete, or in special cases not quite but only approximately complete, equilibrium takes place between the solid and liquid phases (see a separate chapter in the following). And even in the more quickly cooled effusive rocks no completely lacking equilibrium appears, but an imperfect equilibrium, where the degree of imperfection is of a somewhat changeable nature.

With regard to the relations of crystallization, especially in the deep-seated igneous rocks, it is of subordinate importance whether the *simultaneous* crystallization of the final components takes place at a constant temperature (*eutectic point*) or—by a small displacement of the quantitative proportion between the components—within a small interval of temperature for a short distance along a *eutectic boundary curve*. Bowen disputes the justification of the term "gabbroidic eutectic" which I have previously used. I find, however, supported by Bowen's experimental investigations, that this term must be maintained, when we emphasize the fact that here we have to do with a short distance on a curve and not with a point.

We shall now examine the crystallization of the pyroxene minerals and plagioclase, especially in gabbros and norites, and shall commence with the well-known orbicular norite (orbicular quartz-norite) at Romsaas, in the Archaen formation, 50 km. southeast of Kristiania, described by several earlier investigators, especially C. Bugge.[1]

Romsaas, which is a small hill rising about 60 m. above the surrounding gneiss, consists chiefly of quartz-norite, with which

[1] *Kristiania Vidensk. Selskab*, 1906. Here the earlier works of L. Meinich (1878), Th. Hiortdahl (1878), and K. v. Chrustschoff (1897) are cited.

is connected another quite subordinate variety of norite (see the analysis in the chapter on norite in Part II). The entire

MINERALS AND ROCKS FROM ROMSAAS

Percentage of	Hypersthene	Biotite	Plagioclase			Orbicule	Intervening Mass	Orbicular Norite		Common Norite (with Pyrites)
								Calculated	Analyzed	
No.	41	42	43*a*	43*b*	43*c*	44	45	46	47	48
SiO_2	53.3	37.64	52.33	56.25	57.15	51.55	61.28	54.4	52.75	52.86
TiO_2	......	1.57	......	......	......	0.58	0.40	0.5	1.19 }	9.22
Al_2O_3	3.0	20.15	29.99	27.93	27.20	4.45	21.58	9.6	10.29 }	
Fe_2O_3	1.2 }	9.83	0.51	0.45	0.32	0.50	0.22	0.4	0.35	(0.47)
FeO	16.5 }		......	......	......	14.50	1.59	10.6	11.92	11.81
MnO	0.3	0.49	......	......	......	0.50	0.20	0.4	0.46	......
MgO	23.7	16.44	0.97	0.19	0.14	22.08	1.85	16.0	15.61	17.15
CaO	2.0	0.70	11.64	9.59	9.08	2.61	7.51	4.1	4.21	4.58
Na_2O	......	2.57	4.80	5.49	6.01	0.64	4.44	1.7	1.66	1.23
K_2O	......	7.01	0.42	0.16	0.26	0.56	0.74	0.65	0.81	1.03
P_2O_5	......	1.23	......	......	......	0.11	0.52	0.23	0.40	0.53
Ign.	......	1.70	......	0.22	0.27	1.28	0.40	1.0	0.92	0.99
Total	100.0	99.93	100.66	100.28	100.43	99.36	100.77	99.6	100.57	99.87

EXPLANATION

No. 41: Hypersthene (somewhat changed to hornblende) from the orbs, the medium of four well-conformable analyses, by Hiortdahl (2), Meinich, and Chrustschoff.—No. 42: Biotite from the orbicules by Bugge.—No. 43*a-c:* Different analyses of plagioclase from the intervening mass between the orbs, No. *a* (from the kernel of the plagioclase) by Meinich, Nos. *b*, *c* (from the exterior parts of the plagioclase) by Chrustschoff.—No. 44, the orb, and No. 45, the intervening mass between the orbs, by Bugge.—No. 46, the coarse-grained orbicular norite, calculated by myself, and originating from 70 per cent orbs (No. 44) and 30 per cent intervening mass (No. 45).—No. 47 orbicular norite with smaller orbs, analyzed by Bugge.—No. 48, the ordinary norite, analyzed by Bugge, whose analysis applies to a rock with 3.88 per cent pyrrhotite which I have deducted from the analysis, and recalculated to the sum found.

igneous field[1] has an area of only *ca.* 65,000 sq. m. In some places, in part directly at and in part quite close to the boundary of the norite field, the rock has an orbicular structure. We may

[1] See map and profile in Beyschlag, Kreusch, Vogt, *Erzlagerstätten*, I (2d ed., 1914), Fig. 191.

distinguish a variety (see Figs. 7 and 8) with large orbs of a diameter from about 4 to 8 or 9, mostly about 5 cm., and another variety with small orbs, of a diameter of about 2 cm. These orbs consist in the interior exclusively of hypersthene (0.28 $FeSiO_3$:0.72 $MgSiO_3$, if we leave out of consideration the links containing CaO and

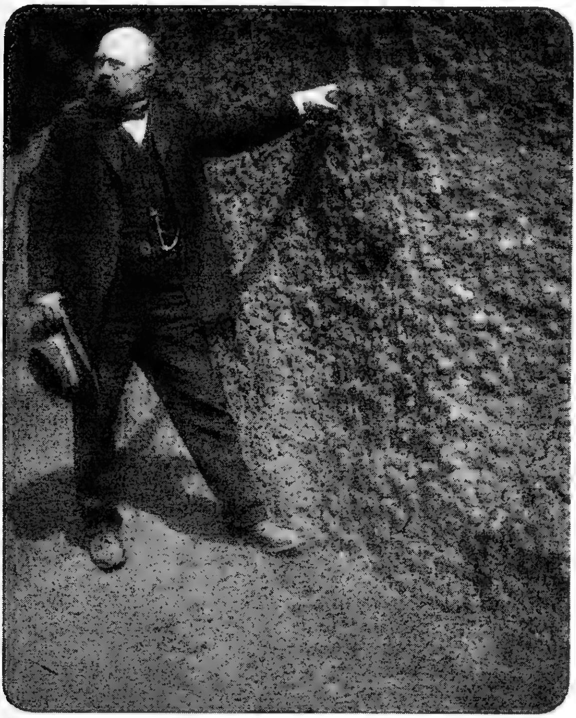

Fig. 7.—Orbicular quartz-norite from Romsaas, Norway

Al_2O_3) which, however, has been changed in part to hornblende. In the exterior part appears also some biotite which partly lies as shells or scales on the outside of the orbs (see Fig. 8). According to the detailed calculation which I effected on the spot, the rock with the large orbs consists (according to weight) of 70 per cent of orb substance (analysis No. 44) and 30 per cent of intervening mass (analysis No. 45). On this basis, the composition, No. 46, of the entire rock is calculated by myself. The close conformity

between analyses Nos. 46–48 prove that the orbicular rock, Nos. 46 and 47, is only a facies near the boundary of the ordinary quartz-norite (No. 48).

The mineralogical composition of the orbs (with some biotite in the exterior part) of the intervening mass (by Bugge), and of the entire quartz-norite (from my own calculation) are as follows:

	Orb	Intervening Mass	Total Quartz-Norite
Hypersthene	92.0	0	63.0
Biotite	7.2	10.0	8.0
Plagioclase	tr.	73.5	23.8
Quartz	....	15.0	4.0
Rutile	0.55	0.39	0.5
Apatite	0.23	1.18	0.9
Total	100.0	100.0	100.0

In addition there sometimes appear in the intervening mass small individuals of garnet, exceptionally also of pyrrhotite ("nickel pyrrhotite") in very small quantity.

The plagioclase of the intervening mass, which on the average may be placed at 47 Ab:3 Or:50 An, therefore almost exactly Ab_1An_1 varies between $Ab_{42}An_{58}$ (with a little Or, analysis No. 43*a*, with extinction 7°15′ on 001 according to Bugge) in the kernel of the individuals and $Ab_{55}An_{45}$ (analysis No. 43*c*, with extinction 3°30′) in the exterior zone. Locally the plagioclase contains still more Ab, according to Bugge with extinction on 001 of 1°15′, corresponding to about $Ab_{62}An_{38}$ (in both cases with a little Or). Concerning this matter we refer to some remarks in a following chapter.

Naturally the orbs are first formed, and only later the intervening mass solidified. Near the center of the orbs, the hypersthene individuals are to a great extent radially arranged. In the exterior part of the orbs[1] we find, on the other hand, an indication of concentric structure (see Fig. 8).

If we leave apatite and rutile out of consideration, we may distinguish between the following stages of crystallization (Fig. 9):

[1] Some inclusions of the material of the intervening mass also appear here and there in the orbs (cf. p. 320).

(1) hypersthene in great quantity (stage i); (2) biotite, in the exterior part of the orbs simultaneously with the crystallization of the rest of the hypersthene (stage ii), then followed some biotite alone, as scales or thin shells on the outside of the orbs (stage iii);

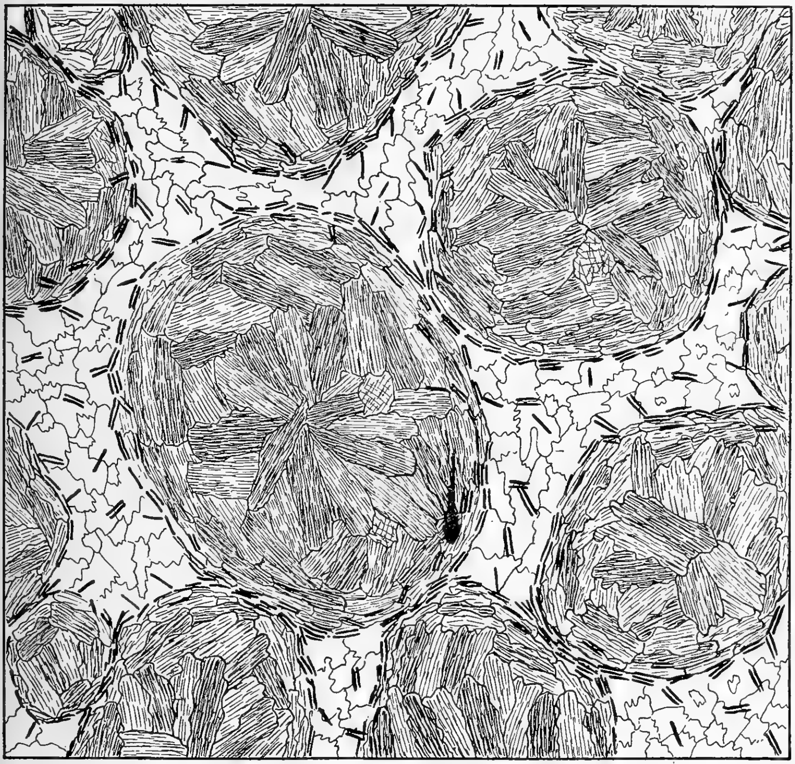

FIG. 8.—Section through the orbicular quartz-norite from Romsaas, Norway. Natural size. Lightly shaded mineral is hypersthene, dark is biotite, and white is feldspar and quartz.

(3) after the orbs were formed, the intervening mass had the composition of a biotite-quartz-diorite (analysis No. 45), and the crystallization became that usual in these rocks, viz., at first biotite (the close of stage iii), then plagioclase (stage iv), and finally also quartz (stage v) solidified.

We especially call attention to the following: In the magma, so extraordinarily rich in ferromagnesian metasilicate, *only hypersthene* crystallized at the beginning. Then the formation of this mineral stopped, as the ferromagnesian silicate still remaining in the magma entered into *biotite*. The change from hypersthene to biotite was probably caused by the quantity of H_2O present in the magma, and this quantity had been relatively enriched by the separation of the large quantity of hypersthene. We shall return to this matter in a following chapter.

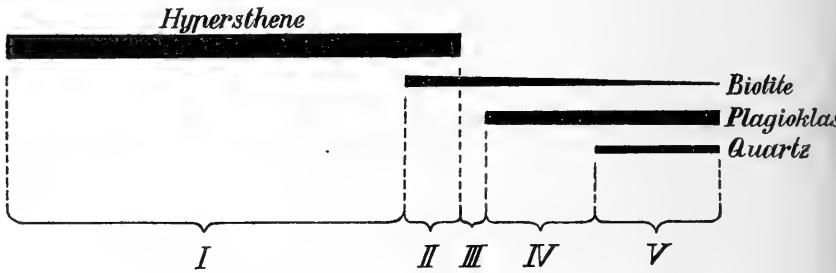

FIG. 9.—Diagram illustrating the different stages of the crystallization of the orbicular quartz-norite from Romsaas.

We further emphasize that from the original magma, so rich in ferromagnesian metasilicate, a quartz-dioritic magma was separated at a far-advanced stage of the solidification and there resulted, by continuous solidification, at last even a magma for special "oligoclase (or andesine) granite dikes," consisting of biotite, andesine ($Ab_{68}An_{32}$), and quartz. We refer to a special chapter in Part II.

The normal quartz-norite (Fig. 10) from Romsaas, consisting of *ca.* 63 per cent hypersthene (included a little secondary hornblende), 8 per cent biotite, 24 per cent plagioclase (Ab_1An_1), 4 per cent quartz, and, in addition, a little apatite, rutile, and pyrrhotite, in part shows accumulation (together-swimming or synneusis structure) of hypersthene individuals which often have a well-developed idiomorphic contour on their boundary toward the plagioclase or quartz. The hypersthene, therefore, must have crystallized completely or in a great measure before the plagioclase and the quartz.

The biotite appears partly in the *exterior* portions of the hypersthene individuals, and partly—and chiefly—grown on to these. The hypersthene in several places shows idiomorphic contours also against the biotite, the latter must accordingly chiefly have been formed at a latter stage than the former. The labradorite

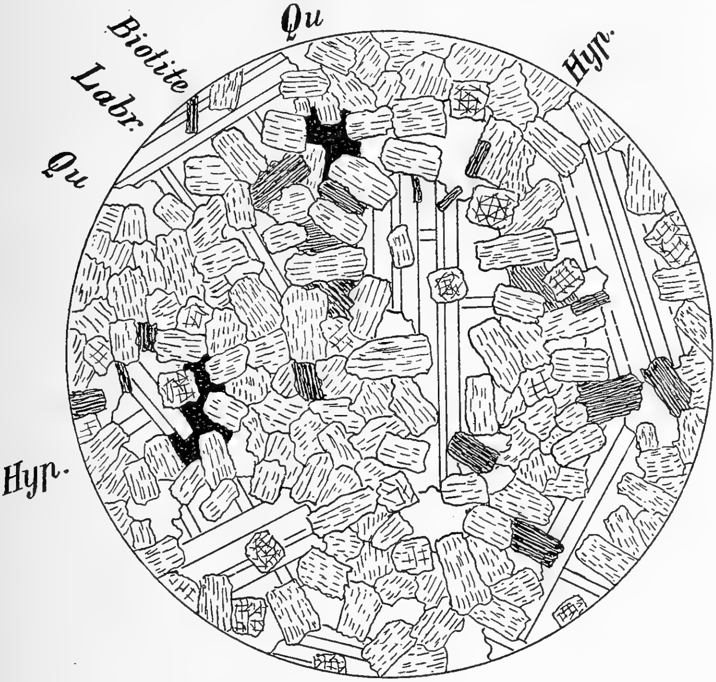

FIG. 10.—Quartz-norite from Romsaas. (Black = pyrrhotite)

and the quartz form an intervening mass between the accumulations of hypersthene, and accordingly crystallized at a somewhat later stage.

The investigation of the Romsaas rocks is in certain respects very instructive, but does not fully inform us of the relation of the crystallization between the hypersthene and the plagioclase, as the ferromagnesian silicate of the later stage entered into biotite instead of into hypersthene. We are therefore going to investigate some rocks where this phenomenon does not appear.

In the norites and gabbros, especially *rich* in orthorhombic or in monoclinic *pyroxene* (hypersthene-norites and diallage-gabbros) with relatively little, say 10, 20, or 25 per cent of labradorite, the pyroxene individuals to a very great extent show an idiomorphic contour against the plagioclase. Further, the pyroxene individuals are often to some extent accumulated in aggregates, consequently showing synneusis structure. On the other hand, the plagioclase shows no signs of idiomorphism but only appears as

FIG. 11.—Photomicrograph (18:1)

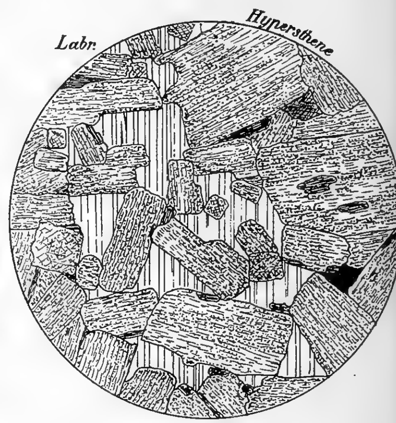

FIG. 12.—Drawing (11:1)

Norite from Meseel, Norway. Hypersthene with predominant idiomorphic outlines against the labradorite. The photograph represents the lower part of the drawing. The shaded mineral in small quantity is hornblende, the black pyrrhotite.

mesostasis (*Zwischenklemmungsmasse*) between the pyroxene individuals. This may be explained by the fact that an essential part of the orthorhombic or monocline pyroxene in question had solidified even before the commencement of the crystallization of the plagioclase. We must not draw the conclusion, however, that the pyroxene individuals in their entirety had crystallized at an earlier stage than the plagioclase. On the contrary, in some of the pyroxene individuals, we find the idiomorphic contour against the plagioclase lacking, and this must indicate a simultaneous crystallization of both minerals during the last stage of the solidification. As an example we refer to the photograph (enlarged 18:1) and

drawing (enlarged 11:1) of a hypersthene-norite from Messel (about 10 km. from Arendal, Norway), containing about 20 per cent labradorite (Ab_1An_1), nearly 80 per cent hypersthene (consisting, according to the determination of the optical character and of the

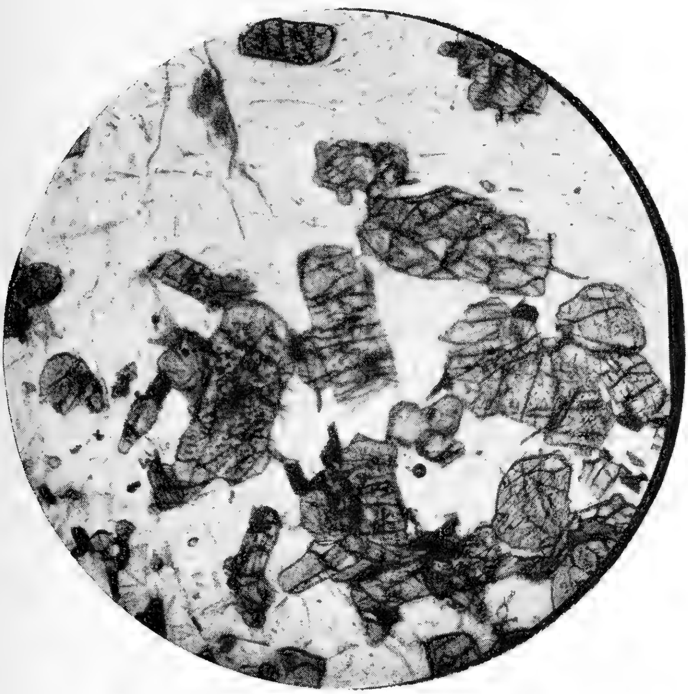

FIG. 13.—Norite from Skougen, Norway. The hypersthene, to a great extent, has idiomorphic outlines against the labradorite, and the hypersthene in several places shows synneusis structure. The biotite is inclosed in the exterior parts of the hypersthene individuals or is grown on these. (24:1.)

optic axial angle, of about $0.25\ FeSiO_3 : 0.75\ MgSiO_3$), a little hornblende, and a little pyrrhotite, but no biotite, diallage, and oxidic iron ore.

In gabbros and norites, containing somewhat more plagioclase (labradorite), say 30, 35, 40, or 45 per cent, and correspondingly less pyroxene, we still find the pyroxene individuals to greater or lesser extent with idiomorphic contours against the plagioclase, while the latter lacks idiomorphism.

As an example we refer to Figures 20 and 21 and to Figure 13, microscopic photograph of a norite from Skougen in Bamle, Norway. This rock consists, according to microscopic investigations supported by chemical analysis (see Part II), of about 47 per cent hypersthene (according to the analysis of the

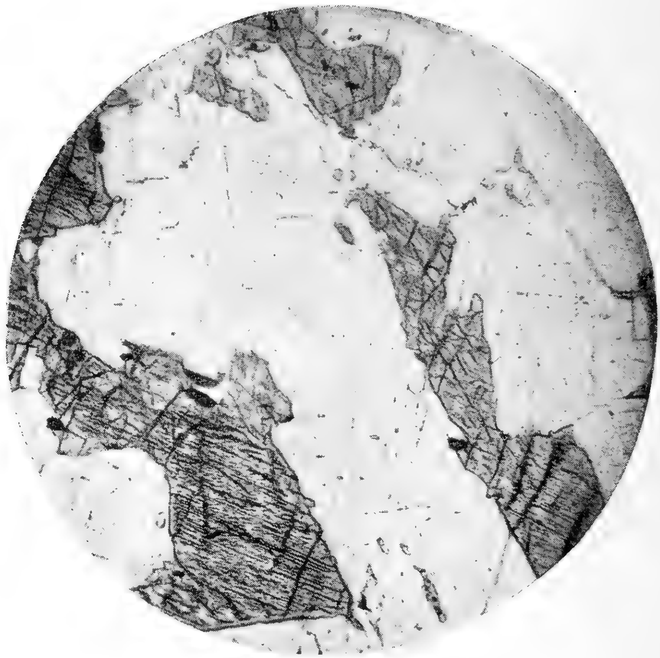

FIG. 14.—Anorthosite from Hitterö, Norway. The labradorite (Ab_1An_1, light) has idiomorphic contours against the hypersthene. (25:1.)

rock and the optical determination calculated as 0.32 $FeSiO_3$: 0.68 $MgSiO_3$), a trifle secondary hornblende, 48 per cent labradorite (*ca.* 38 Ab, 4 Or, 58 An, or about Ab_2An_3), 3 per cent biotite, and 1–2 per cent magnetite-ilmenite, see later (Figs. 31–32), 0.07 per cent apatite, and a little pyrite (0.24 per cent S).

On the other hand, in rocks especially *rich in plagioclase* we find throughout the idiomorphism more or less well developed by the plagioclase, but not by the pyroxene. This applies to all anorthosites which I have investigated, where the ferromagnesian

silicates, indifferently whether hypersthene (Fig. 14), diallage, or olivine (see Figs. 23 and 24), for an essential part form a mesostasis between the plagioclase (labradorite or sometimes bytownite).

The question in hand, I have to some extent considered in a paper, accompanied by several analyses, published in the *Quart. Jour. of the Geological Society*, 1909, on labradorite-norite with porphyritic labradorite crystals from Flakstadöen in Lofoten. Referring to the quantitative analysis of this rock, given in the chapter on anorthosite-norite in Part II, I shall here give a short résumé. The entire rock consists of:

ca. 70.65 per cent labradorite, 55 An, 39 Ab, 6 Or
ca. 6.3 per cent Fe_3O_4 } 7.2 per cent "titanomagnetite"
ca. 0.9 per cent $FeTiO_3$ }
ca. 10.0 per cent hypersthene
ca. 10.0 per cent diallage
ca. 2.3 per cent biotite
ca. 0.9 per cent apatite

Relatively to the entire rock, there first crystallized 23 per cent prophyritic labradorite of a composition 61 An, 33 Ab, 6 Or, and occurring as very large, up to 15–18 cm. long and 6–8 cm. wide, crystals, thick tabular along (010). The remainder, 77 per cent, form a coarse-grained groundmass, consisting of:

ca. 61.9 per cent labradorite, 52 An, 42 Ab, 6 Or
ca. 8.1 per cent Fe_3O_4 } 9 per cent "titanomagnetite"
ca. 0.9 per cent $FeTiO_3$ }
ca. 13.0 per cent hypersthene
ca. 13.0 per cent diallage
ca. 3.0 per cent biotite
ca. 0.12 per cent apatite

In this groundmass the labradorite continued crystallizing, and some magnetite (or "titanomagnetite") commenced to form, of which more below. Only at a somewhat later stage, after the quantity of hypersthene and diallage had risen somewhat above 30 per cent, the pyroxene commenced forming.

The *hyperitic* (or ophitic) texture of plagioclase crystals, with tabular development along (010), appears in the gabbro rocks only when the latter contain at least about 55 per cent plagioclase (labradorite). The laths of plagioclase show partial idiomorphism

against the hypersthene (see, for example, Figs. 15 and 16) or the diallage, which ordinarily entirely lack idiomorphism. We may consequently draw the conclusion that the crystallization of labradorite in these plagioclase-rich rocks must have commenced before the beginning of the solidification of the pyroxene. The idiomorphism of the labradorite, however, is often only quite slightly developed, as illustrated in Figures 15 and 16. This tells us that

FIG. 15.—Photomicrographic (19:1)

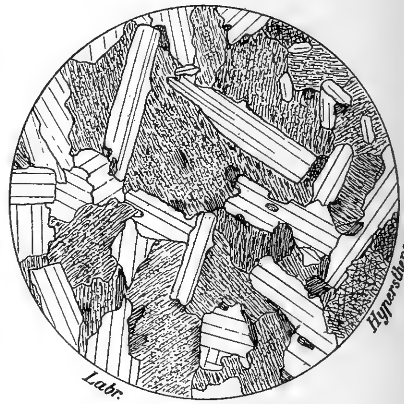

FIG. 16.—Drawing (35:1)

Hyperitic- (or ophitic-) structured norite from Erteli, Norway. Consists of about 56 per cent labradorite (*ca.* Ab_3An_7), 41 per cent hypersthene (0.31 $FeSiO_3$: 0.69 Mg SiO_3), a little magnetite and pyrrhotite (0.07 per cent S, see Fig. 46), 0.09 per cent apatite; see analysis in Part II. The drawing (35:1) represents the central and lower parts of the photograph (19:1).

only a certain small part of the labradorite, in this case, had solidified before the pyroxene began forming.

Rosenbusch[1] emphasized, and rightly, with regard to the gabbros, that in rocks *rich in plagioclase*, the plagioclase, and in the varieties *rich in diallage*, the diallage, develops in idiomorphic individuals, and further, that in rocks rich in diallage, the idiomorphism of the diallage is the more prominent the greater its quantity. "Man wird also scheinbar genötigt ein gewisses Schwanken in der Reihenfolge der Ausscheidungen anzunehmen." But Rosenbusch did not engage in the physicochemical interpretation of the phenomenon.

[1] *Mikroskop. Phys. d. Mass. Gest.* (4th ed., 1907), II, 1, p. 364.

According to my own extensive investigations of the question in hand (whereof I have only given a very short résumé) the facts previously set forth by Rosenbusch are confirmed, and we draw the conclusion that *the sequence of crystallization of plagioclase and pyroxene depends upon the relative quantity* of the two minerals, for *the crystallization commences with the solidification of the mineral present in excess of a certain limit.*

I have bestowed much labor on the determination of this limit, trying to get it as exactly as possible, by structural investigation on a series of norites, and partly also of gabbros, of which we have numerous quantitative analyses, so that the proportion by weight between the plagioclase (with a determined Ab:An relation) and the pyroxene may be quite exactly calculated. We then find that the limit essentially depends on the composition of the plagioclase.

As an example may be mentioned that in a diallage-bearing quartz-norite with about 58 per cent SiO_2 (and standing on the boundary near quartz-hypersthene diorite), containing about 28 per cent pyroxene (hypersthene with a little diallage), about 60 per cent plagioclase (andesine, $Ab_{68}An_{40}$), a trifle orthoclase, about 4 per cent biotite, about 1 per cent iron ore, and about 5 per cent quartz, the hypersthene in a great measure appears with idiomorphic contour against the plagioclase (andesine). In rocks with basic labradorite, as Ab_1An_2, the plagioclase on the other hand partly shows an idiomorphic contour against the pyroxene, even when there is as much pyroxene as 35–40 per cent present. The individualization boundary, determined by the sequence of crystallization, amounts approximately in the deep-seated rocks to:

With $Ab_{30}An_{70}$ about	45–50 per cent pyroxene	
With $Ab_{40}An_{60}$ about	40–45 per cent pyroxene	
With $Ab_{50}An_{50}$ about	35–40 per cent pyroxene	the remainder plagioclase
With $Ab_{60}An_{40}$ about	25–30 per cent pyroxene	
With $Ab_{70}An_{30}$ probably	15–20 per cent pyroxene	

By pyroxene we here understand, with regard to the norites, hypersthene of the common composition of these rocks, viz., 0.3–0.35 $FeSiO_3$:0.7–0.65 $MgSiO_3$, and, with regard to the gabbros, diallage with about the corresponding iron content.

We are able to obtain a more detailed determination of the individualization boundary by studying a series of porphyritic rocks, with phenocrysts of plagioclase when this mineral is in excess, and of pyroxene when the ferromagnesian silicate is in excess. I am, however, far from having sufficient material for such a precision-determination.

Because of physicochemical considerations we must draw the conclusion that the *individualization boundary* here shown is a *eutectic* boundary between pyroxene and plagioclase. And our individualization boundary shows almost exactly the same course as the eutectic boundary curve between diopside and Ab+An, determined by Bowen at a pressure of one atmosphere.

If we imagine a quartary system, consisting of Ab, An, $CaMgSi_2O_6$ (Diops), and $CaFeSi_2O_6$ (Hed), this system will separate into two fields, each consisting of a continuous mix-crystal. Ab+An surely, and Diops+Hed almost certainly, belong to type I. The difference between this quartary system and the ternary system Ab+An:Diops (Fig. 6), or Ab+An:Qu (Fig. 5), is chiefly that for the melting-plane of the independent component in the ternary system (ex. Qu on Fig. 5) is substituted a melting-plane of a binary mix-crystal combination (Diops+Hed).

The eutectic boundary between basic plagioclase and hypersthene I term the "noritic," between basic plagioclase (labradorite, exceptionally bytownite) and monoclinic pyroxene (diallage) the "gabbroidic," and between somewhat more acid plagioclase (andesine, in some cases oligoclase) and the ferromagnesian constituent, the "dioritic" eutectic boundary. We shall calculate the composition of this boundary on the assumption that it has the following course with varying proportions of Ab and An:

45 per cent hypersthene	(or diallage):	55 per cent $Ab_{20}An_{80}$
40 per cent hypersthene	(or diallage):	60 per cent $Ab_{35}An_{65}$
35 per cent hypersthene	(or diallage):	65 per cent $Ab_{50}An_{50}$
33 per cent Augite	(or diallage):	67 per cent $Ab_{55}An_{45}$
27 per cent Augite	(or diallage):	73 per cent $Ab_{60}An_{40}$
20 per cent Augite	(or diallage):	80 per cent $Ab_{75}An_{25}$

For hypersthene we assume: 48 $MgSiO_3$, 30 $FeSiO_3$, 5 $MgAl_2SiO_6$, 4 $MgFe_2SiO_6$, 13 $CaFeSi_2O_6$ = 51.2 per cent SiO_2, 2.5 Al_2O_3, 2.5 Fe_2O_3, 2.9 CaO, 20.1 FeO, 20.8 MgO. And for the plagioclase respectively $An_{78}Ab_{18}Or_4$, $An_{62}Ab_{32}Or_6$, and $An_{47}Ab_{47}Or_6$.

	Noritic Eutectic Line between Plagioclase and Hypersthene		
	I	II	III
	45 Hyp: 55 $An_{78}Ab_{18}Or_4$	40 Hyp: 60 $An_{62}Ab_{32}Or_6$	35 Hyp: 65 $An_{47}Ab_{47}Or_6$
SiO_2	49.7	51.9	54.5
Al_2O_3	19.3	19.1	18.8
Fe_2O_3	1.1	1.0	0.9
FeO	9.1	8.1	7.1
MgO	9.3	8.3	7.3
CaO	9.9	8.7	7.1
Na_2O	1.2	2.3	3.6
K_2O	0.4	0.6	0.7

For the diallage of the gabbros we assume: 40 $CaMgSi_2O_6$, 36 $CaFeSi_2O_6$ 7 $MgAl_2SiO_6$, 11 $FeFe_2SiO_6$, 6 $MgSiO_3$=47.5 per cent SiO_2, 3.8 Al_2O_3, 6 Fe_2O_3, 13.1 FeO, 11.1 MgO, 18.5 CaO. And for the augite of the diorite: 52 $CaMgSi_2O_6$, 34 $CaFeSi_2O_6$, 3 $MgAl_2SiO_6$, 6Fe Fe_2SiO_6, 5 Mg SiO_3=50.45 per cent SiO_2, 1.65 Al_2O_3, 3.3 Fe_2O_3, 11.3 FeO, 12.2 MgO, 21.1 CaO.

	Gabbroidic Eutectic Line between Plagioclase and Diallage			Dioritic Eutectic Line between Plagioclase and Augite		
	I*b*	II*b*	III*b*	IV	V	VI
	45 Diallage: 55 $An_{78}Ab_{18}Or_4$	40 Diallage: 60 $An_{62}Ab_{32}Or_6$	35 Diallage: 65 $An_{47}Ab_{47}Or_6$	33 Augite: 67 $An_{42}Ab_{48}Or_{10}$	27 Augite: 73 $An_{32}Ab_{56}Or_{12}$	20 Augite: 80 $An_{22}Ab_{66}Or_{12}$
SiO_2	48.0	50.5	53.2	55.1	57.5	60.2
Al_2O_3	19.9	19.7	19.3	18.5	18.5	18.8
Fe_2O_3	2.7	2.4	2.1	1.1	1.0	0.7
FeO	5.9	5.2	4.6	3.7	3.0	2.3
MgO	5.0	4.4	3.9	4.0	3.3	2.4
CaO	16.9	14.9	12.6	12.6	10.4	7.8
Na_2O	1.2	3.3	3.6	3.8	4.8	6.2
K_2O	0.4	0.6	0.7	1.1	1.5	1.6

In the anchi-eutectic norites or gabbros with rather basic plagioclase (Ab_1An_2), the simultaneous crystallization of plagioclase and pyroxene will commence at about II (or II*b*) and be finished at about III (or III*b*), or at still less pyroxene and more of a relatively Ab-rich plagioclase, about equal to IV. And in the medium-basic anchi-eutectic diorites the simultaneous crystallization of plagioclase and pyroxene will commence about at IV (or between IV and V) and finish at about VI.

[*To be continued*]

TYPES OF ROCKY MOUNTAIN STRUCTURE IN SOUTHEASTERN IDAHO[1]

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INTRODUCTION

Since 1909 the United States Geological Survey has been making detailed studies of portions of the western phosphate field, chiefly in southeastern Idaho. This region contains a series of sedimentary rocks 40,000 feet or more thick, including large bodies of high-grade phosphate rock that will prove of great economic importance for the future, if not for the present. There is interesting geologic structure and a variety of problems covering a wide range of geologic and geographic phenomena.

[1] Read before the Geological Society of America, December 30, 1919; published by permission of the Director of the U.S. Geological Survey.

The area included in the detailed surveys is nearly 3,000 square miles comprised in the Fort Hall Indian Reservation, and in the Montpelier, Slug Creek, Crow Creek, Lanes Creek, Freedom,

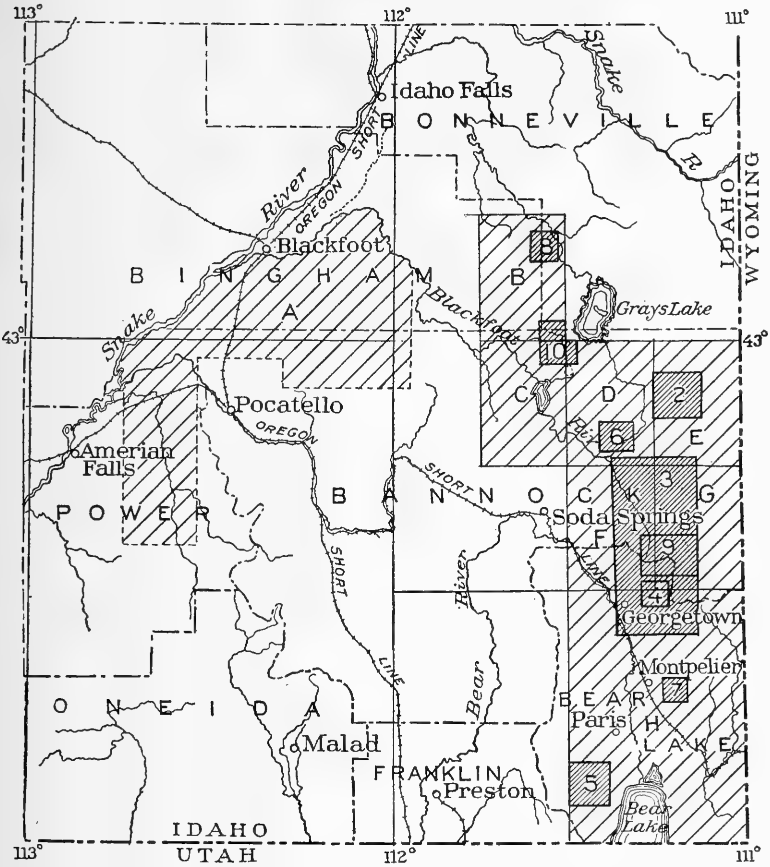


FIG. 1.—Index map of southeastern Idaho: *A*, Fort Hall Indian Reservation; *B-H*, seven quadrangles; *B*, Cranes Flat; *C*, Henry; *D*, Lanes Creek; *E*, Freedom; *F*, Slug Creek; *G*, Crow Creek; *H*, Montpelier. The areas lettered 2–10 are those illustrated by the corresponding figures.

Henry, and Cranes Flat quadrangles, all of which are fifteen-minute quadrangles except the Montpelier, which is a thirty-minute quadrangle. The location of these areas is shown on the accompanying map, Figure 1.

Three semi-detailed reports[1] and a number of shorter papers have already been published and a fourth report[2] is now in press. An additional, more extended report is well advanced in preparation and includes a discussion of the geography, geology, and mineral resources of the seven quadrangles named. The purpose of this paper is to present in advance of the detailed report some of the striking structural types of the region and to discuss briefly certain conditions that attended the development of these structures. The maps used in illustration of the structural features are extracted from the detailed geologic maps of the quadrangles mentioned. Their locations are shown on the index map, Figure 1.

GENERAL STRUCTURAL FEATURES

The stratigraphic series in southeastern Idaho includes more than sixteen recognized unconformities. Most of them do not appear to record great crustal disturbances, but a few indicate changes of considerable magnitude. Several are very striking, both as seen in the field and in cartographic representation

The region is traversed by many folds, some of which exceed 50 miles in length. The more important folds are synclinoria with relatively narrower intervening anticlines or anticlinoria, usually unsymmetrical and inclined or even overturned eastward or northeastward. The axes for long distances are nearly horizontal or slightly undulatory, due to the presence of relatively broad and low transverse folds, and the pitch is gentle, generally toward the north or northwest. The trend of the folds is convex toward the northeast, bending from a little east of north in the Montpelier quadrangle to northwest in the Lanes Creek quadrangle and beyond. This arrangement gives rise to long nearly parallel folds

[1] See especially H. S. Gale and R. W. Richards, "Preliminary Report on the Phosphate Deposits in Southeastern Idaho and Adjacent Parts of Wyoming and Utah," *U.S. Geol. Survey Bull. 430* (1910), pp. 457–535; R. W. Richards and G. R. Mansfield, "Preliminary Report on a Portion of the Idaho Phosphate Reserve," *U.S. Geol. Survey Bull. 470* (1911), pp. 371–451; R. W. Richards and G. R. Mansfield, "Geology of the Phosphate Deposits Northeast of Georgetown, Idaho," *U.S. Geol. Survey Bull. 577*, 1914.

[2] G. R. Mansfield, "The Geography, Geology and Mineral Resources of the Fort Hall Indian Reservation, Idaho, with a Chapter on Water Resources, by W. B. Heroy," *U.S. Geol. Survey Bull. 713.*

somewhat similar to those of the southern Appalachians. The Idaho folds, however, appear to be less regular in form than those of the Appalachian region. The intensity of the folding may be judged by the fact that within the region of the seven quadrangles mapped there are forty-two folds or groups of folds that have been considered of sufficient importance to receive names and to merit individual treatment in a detailed description of the region.

The influence of the transverse folds is seen chiefly in the widening or constriction of the longitudinal folds, in the production here and there of canoe- or cigar-shaped folds, and in the zigzag outcrop of certain formations, which cross the axes of the longitudinal folds.

The principal faults of the region are reverse and are doubtless chiefly associated with the Bannock overthrust, which has a length probably greater than 270 miles and a horizontal displacement certainly not less than 12 miles and perhaps greater than 35 miles. Normal faults are numerous and have produced a wide range of effects upon the pre-existing structures. Possibly some of the faults now regarded as reverse may prove to be normal. The intensity of the faulting is suggested by the fact that about sixty faults or groups of faults are sufficiently noteworthy to receive individual consideration in a detailed description of the region. About half of these are thrusts associated with the Bannock overthrust.

SPECIAL STRUCTURAL FEATURES

The structures to which attention is especially directed in this paper are (1) noteworthy unconformities; (2) "swallowtail" folds; (3) the Bannock overthrust; (4) the Blackfoot fault; (5) drag folds; (6) fan folds; and (7) the Meadow Creek graben. These will be described in the order named.

Noteworthy unconformities.—A very marked unconformity occurs in the southeastern part of the Montpelier quadrangle, where strongly folded Triassic and Jurassic beds pass beneath gently folded or nearly horizontal beds of the Wasatch formation (Eocene). This unconformity is the most striking of all the unconformities of the region. It represents at least the great post-Cretaceous mountain-building epoch of the northern Rocky

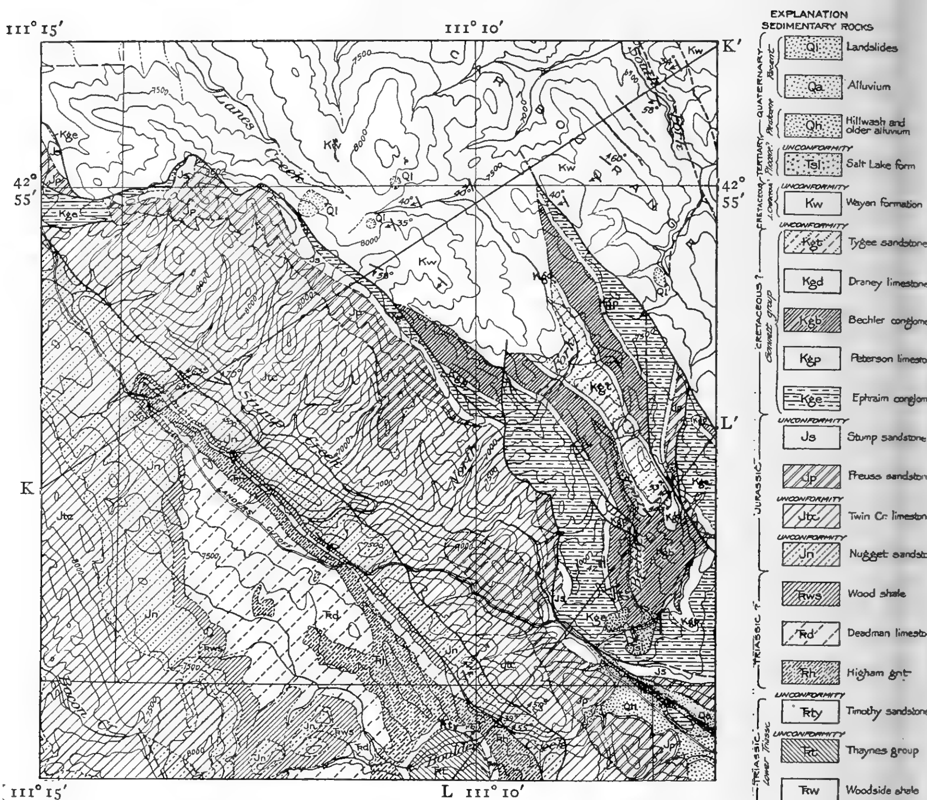


FIG. 2 *A*.—Map of part of the Caribou Range, Freedom quadrangle, showing the fault zone of the Bannock overthrust and the unconformity between the Wayan formation and the Gannett group.

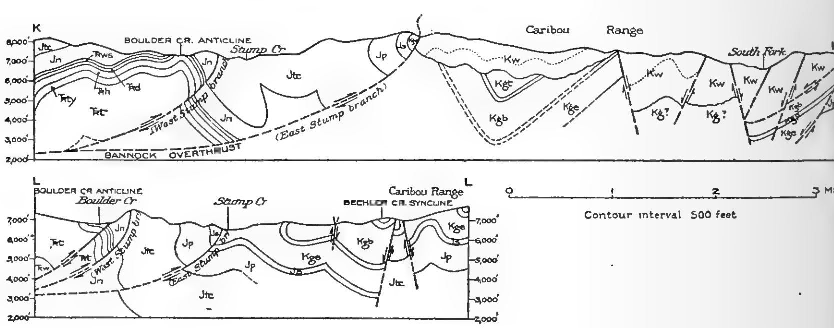


FIG. 2 *B*.—Sections along the lines *K–K'* and *L–L'* of Fig. 2 *A*

Mountains and a succeeding epoch of erosian long enough to remove all Eocene beds earlier than the Wasatch, if such were ever deposited, the Cretaceous formations, and two of the upper Jurassic formations.

A pronounced unconformity occurs between the Wayan formation and the underlying Gannett group, both of supposed Lower Cretaceous age. Figure 2 shows one of the localities where this unconformity appears. A syncline composed of members of the Gannett group is overlapped on the northwest by the Wayan formation, composed of beds folded in a manner comparable to that of the Gannett group but not yet differentiated into members. Another exposure of the same unconformity, perhaps even more striking, occurs about 3 miles east of the locality shown on the map.

"Swallowtail" folds.—These are folds in which the axes are nearly horizontal but are affected by cross folds in such manner that the outcropping formations as represented on the map resemble a swallow's tail. A remarkable group of folds having this form occurs in the Slug Creek and Crow Creek quadrangles, as shown in Figure 3, and has great economic importance because of its contained beds of high-grade phosphate. The group lies in the arc of curvature of the trend lines previously mentioned. A transverse syncline near the northern part of the area shown on the map causes the widening of the two lateral synclines and depresses the axis of the intervening anticline. It also causes the widening of the canoe-shaped fold partly shown in the northwestern part of the area. Other transverse axes cause the ending of the canoe-shaped fold and the changes in width of the middle and southern portions of the "swallowtail" folds. The axes of the transverse folds are roughly radial to the curvature of trend of the "swallowtail" folds.

The Bannock overthrust.—This great overthrust was first described in 1912.[1] Since that date the writer has had opportunity to extend his observations along the fault line and to secure new

[1] R. W. Richards and G. R. Mansfield, "The Bannock Overthrust: a Major Fault in Southeastern Idaho and Northeastern Utah," *Jour. Geol.*, Vol. XX (1912), pp. 681–707.

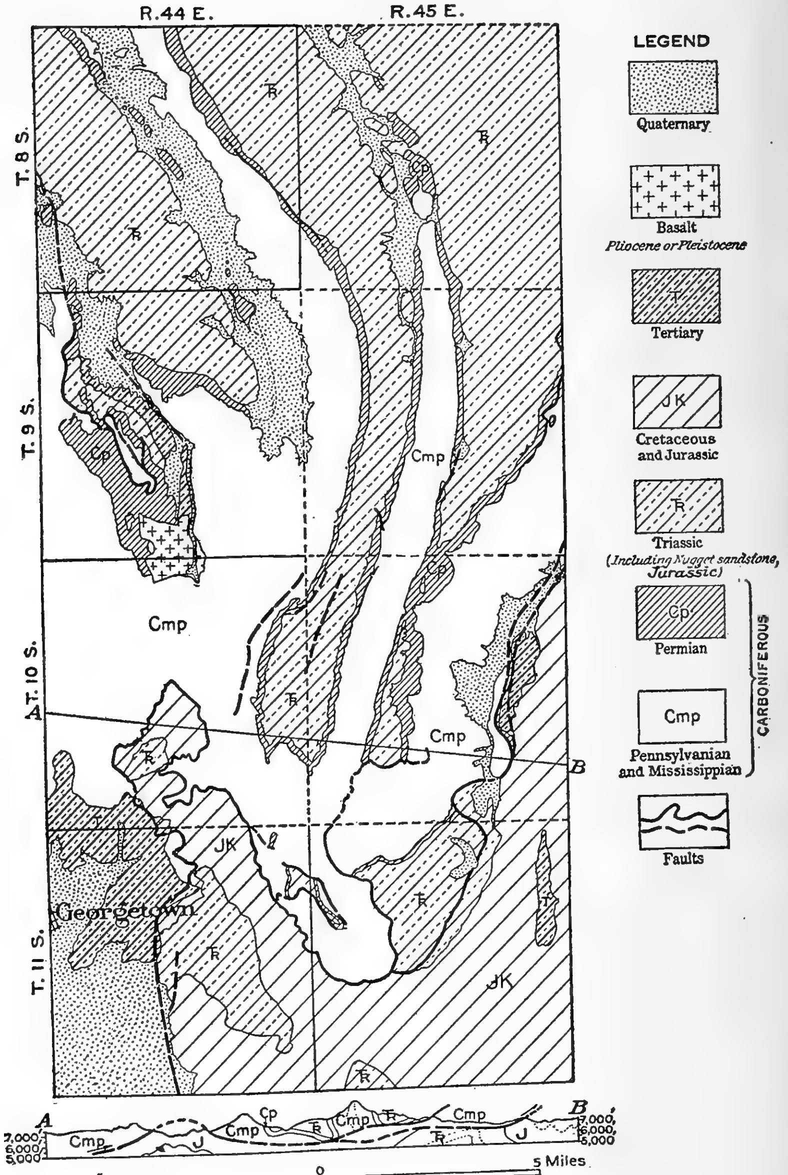


Fig. 3.—Map with geologic structure section of "swallowtail" folds in adjacent parts of the Slug Creek, Crow Creek, and Montpelier quadrangles, showing the folding and erosion of the plane of the Bannock overthrust, U.S.G.S.

data. The folding of the fault plane, previously announced, is well shown in Figure 3, both in plan and in section. Two aspects of this folding should be emphasized: (1) the folds of the fault plane are of a simpler and more open type than are those of the upper or lower fault blocks, as shown in the geologic structure section; (2) the fault plane previous to its deformation must have been nearly horizontal, else moderate folding would not have raised it to the level of erosion.

The underlying block is in most places composed of Mesozoic rocks more or less intensely folded, whereas the upper block is more largely composed of Paleozoic rocks that in the main are more competent strata. Mesozoic rocks, however, form part of the upper block at many localities. The relation of upper to lower block is shown in Figure 4, which represents the faulted area on the boundary line between T. 11 S., R. 44 E., and T. 11 S., R. 45 E., Figure 3. The fault there shown is regarded as a branch of the Bannock overthrust and is so represented on the map, but it may be a window in the main fault plane, as is supposed in the case of the area surrounded by a fault in T. 9 S., R. 44 E.

The Bannock overthrust is in some places a single fault plane, as in part of the area shown in Figure 3, but in other places it becomes complex and is really a fault zone composed of a number of rock slices separated by faults. Several branches are shown in Figure 3, and the branch that separates the Carboniferous from the Triassic rocks in T. 11 S., R. 45 E., is locally overturned and dips eastward. Three branches of the overthrust are shown in Figure 2. The East Stump branch, where crossed by the line of section KK', is practically vertical, but farther northwest it is overturned and dips northeastward. The West Stump branch, where crossed by the lines of sections KK' and LL', is also practically vertical, but at Boulder Creek it dips northeastward.

Figure 5 shows a complex portion of the fault zone near St. Charles, west of Bear Lake. There are probably no less than six faults which divide the rocks into roughly parallel slices east of the belt of Brigham quartzite, which is the easternmost formation of the upper fault block. The trace of the west branch of the fault in this district and southward lies east of a series of

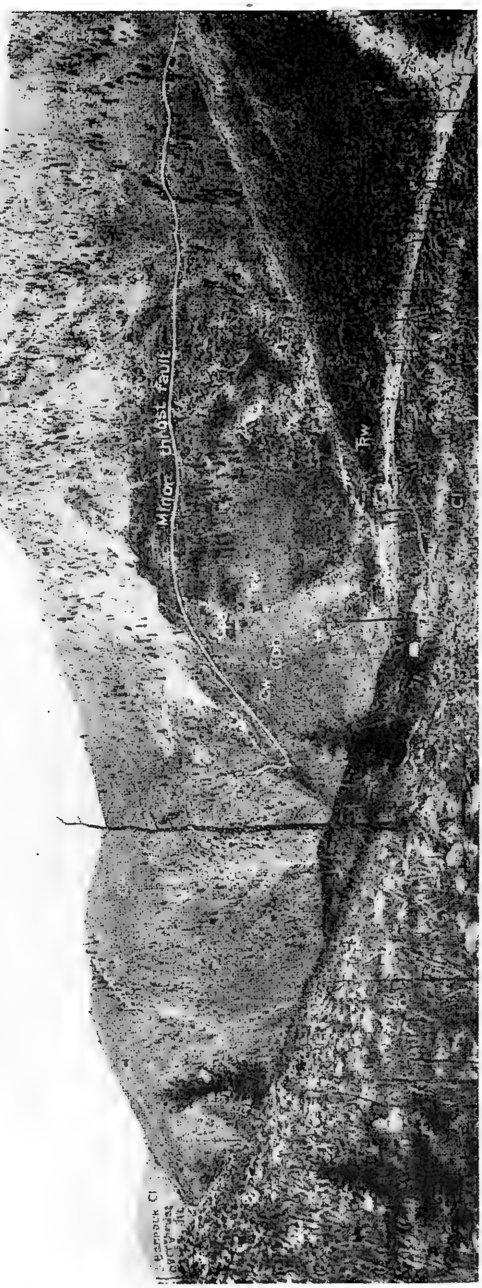


FIG. 4.—Meade Peak and South Canyon, from ridge in center of sec. 12, T. 11 S., R. 44 E. *C*1, Mississippian limestone; *Cw*, Wells formation; *Cpb*, phosphatic shales, etc. (Phosphoria formation); *Cpa*, Rex chert member of Phosphoria formation; *Trw*, Woodside shale; *Jtc*, Twin Creek limestone, U.S.G.S.

topographic sags and laps up on the west side of the adjoining hills. The upper fault block is composed of Cambrian and Ordovician formations, comprising the east limb of a syncline. The rock slices in the fault zone include at least a broken syncline of Ordovician

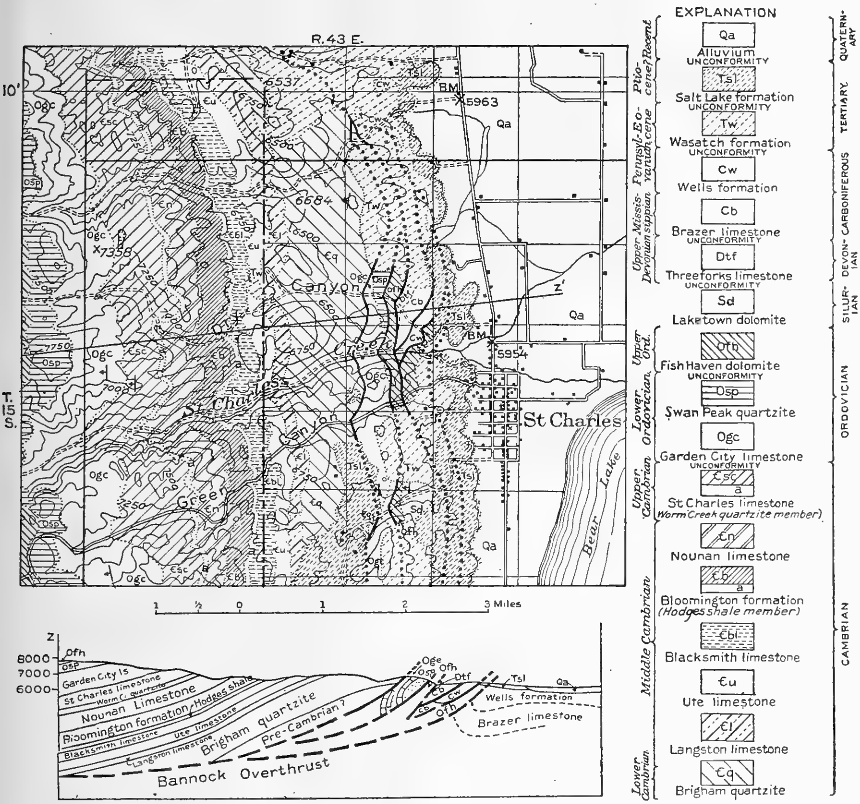


FIG. 5.—Map with geologic structure section of part of the Bear River Range, Montpelier quadrangle, showing the complex fault zone of the Bannock overthrust.

formations and parts of other folds containing Devonian and Carboniferous formations. The easternmost rock slice of the zone is probably composed of Fish Haven dolomite (Upper Ordovician). The structures east of the fault zone, along the line of structure section *ZZ′*, are concealed by Tertiary beds and alluvium, but farther north scattering outcrops of the Wells formation (Pennsylvanian) occur east of this zone. It is thought, therefore, that the

synclinal structure east of the overthrust in the Bloomington-Paris district about 4 miles to the north may continue southward beneath

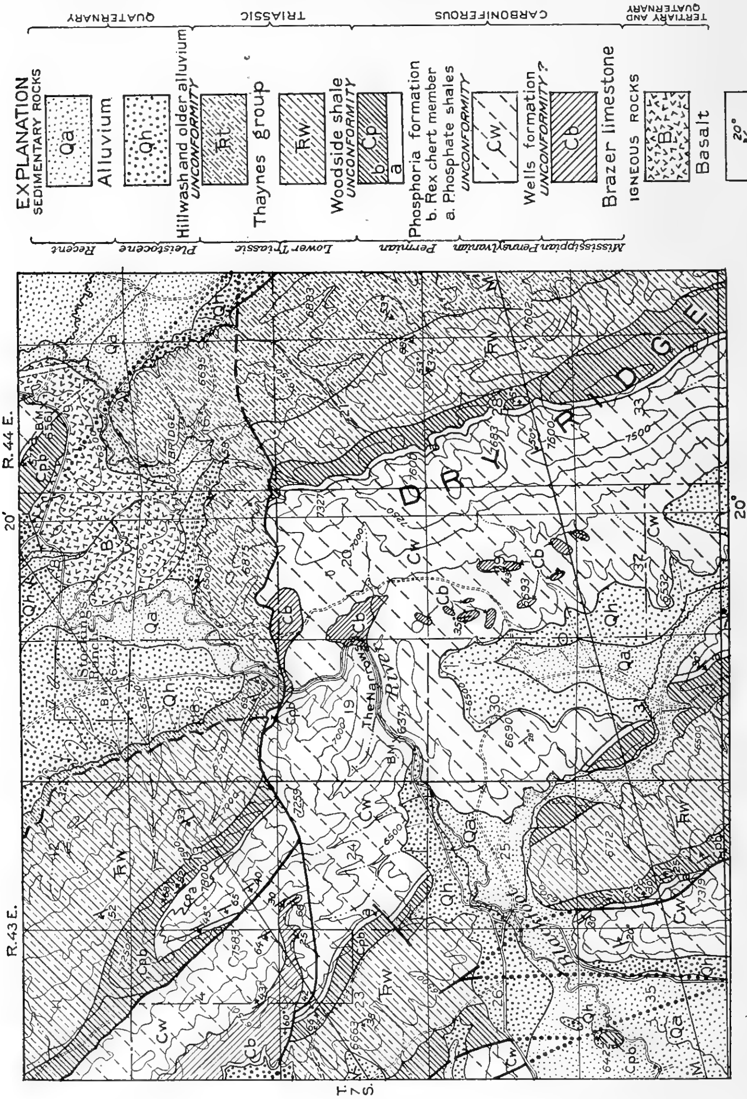


FIG. 6 *A*.—Map of part of the Preuss Range, Lanes Creek quadrangle, showing the Blackfoot fault and associated features.

cover. The dip of the thrust plane as a whole in the St. Charles district is probably gentle. A measurement in Paris Canyon about 6 miles north shows the dip at that locality to be 23°.

The Blackfoot fault.—This fault, which is illustrated in part in Figure 6, takes its name from the Blackfoot River, which it crosses at the upper entrance to the Narrows. The relations of the fault at this point indicate that its plane dips about 33° south. It is

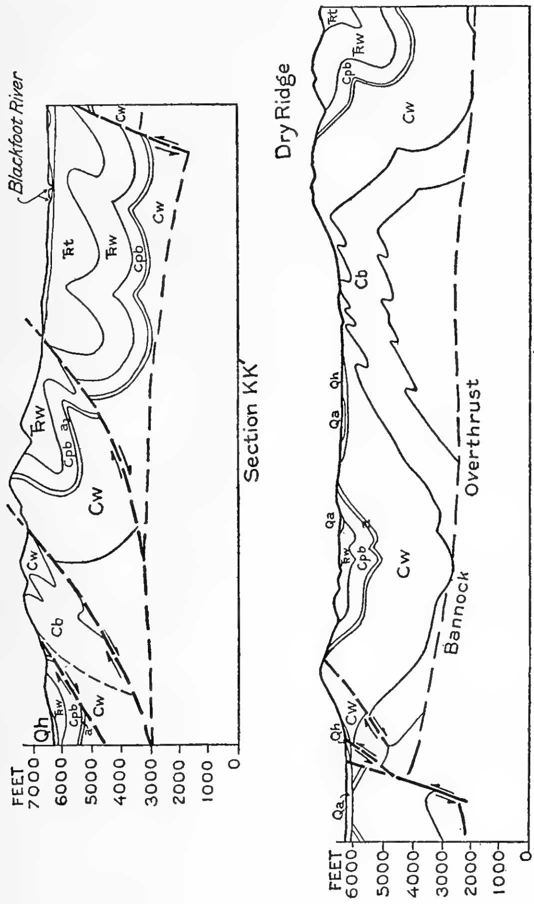


FIG. 6*B*.—Sections along the lines *K–K'* and *M–M'* of FIG. 6*A*

regarded as a thrust fault and is supposed to have originated in a transverse anticline, located near the line along which occurred the maximum yielding of the rocks to the compressive earth stresses of the region. The anticline broke and the southeast limb, which became the upper fault block, swung northward about a pivot located in the vicinity of Timothy Creek in the Freedom

quadrangle, about 5 miles east of the area illustrated. The rock formations cross this creek without apparent displacement by the fault, but they make a pronounced bend, which is favorably located to mark the position of the unbroken portion of the axial zone of the anticline.

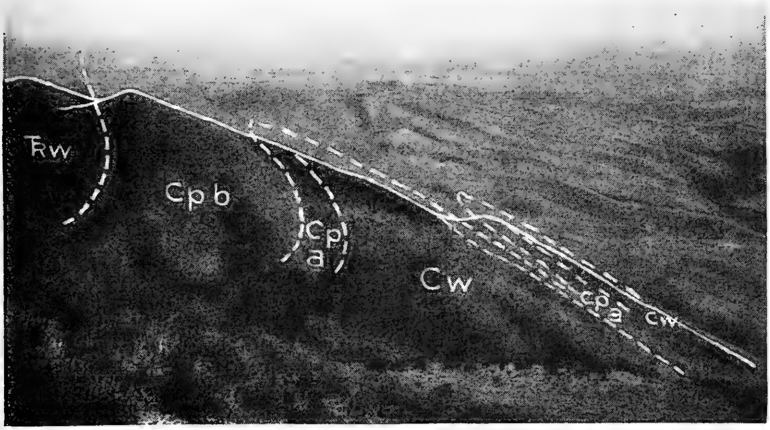
The Blackfoot fault has a known length of about 13 miles westward from the point of origin above indicated. At this distance it disappears beneath basalt. The variations in dips on the flanks of the big anticline in the upper thrust block, some of the strata being locally overturned, and the presence of minor folds, make it difficult to determine the throw of the fault. The maximum observed effect is produced where it cuts the big anticline. The fault is probably offset beneath cover by one of the normal faults shown in the southwestern part of the area.

The upper fault block affords an unusually fine example of the manner in which the outcrops of a formation, such as the Phosphoria, occurring on opposite limbs of an anticline, are spread by the uplift and erosion of a fault block, in which the anticline is included. In the lower block to the north the corresponding outcrops are much nearer together.

The structure section along the line KK' crosses the Blackfoot fault. It shows the anticlinorial and synclinorial character of the folds, their overturning toward the northeast, and the manner in which some of the subsidiary thrusts are thought to have originated. The thrusts are presumed to pass into the Bannock overthrust, which underlies this district.

Drag folds.—Folds of this type, usually sharp and unsymmetrical, occur at a number of places in the region. Several of them are shown in Figure 6 in connection with the big anticline in the upper fault block. One of them is crossed by the line of structure section MM'. Here upper beds of the Brazer limestone (Mississippian) in small sharp folds are locally exposed by the erosion of the overlying Wells formation.

Drag folds on a somewhat larger scale occur on the hills south of Montpelier Canyon about 3 miles east of Montpelier. In Figure 7*A*, a view south along the west flank of Waterloo hill shows an anticline overturned eastward in which curving beds of



A

B

Fig. 7.—*A*, View south along west flank of Waterloo Hill about 3 miles east of Montpelier, Idaho, showing drag folds accompanying large unsymmetrical anticline; *B*, view of same folds northward from different viewpoint, showing west flank of the same anticline. *Cw*, Wells formation, Pennsylvanian; *Cpa*, phosphatic shale member of Phosphoria formation, Permian; *Cpb* Rex chert member of the Phosphoria formation, Permian; *Trw*, Woodside shale, Lower Triassic.

the Rex chert member of the Phosphoria formation (Permian) form the east limb of the anticline and indicate the occurrence of a syncline farther east. In the middle of the view is a sharp anticlinal fold inclined eastward. Near the base of the slope at the right (west) is another drag fold, very sharp and inclined eastward. These drag folds are composed of upper beds of the Wells formation and of the overlying phosphatic shales of the Phosphoria formation, which are relatively less competent than the Rex chert above or the bulk of the Wells formation below. The effect of the drag folds is to duplicate the outcropping beds of phosphatic shales, which appear as separate bands on the hillside. The anticline, of which the drag folds form a part, is largely eroded, but its west limb is exposed in a branch canyon to the north. Figure 7*B* is a view north from a somewhat different viewpoint. It shows the westerly dipping beds of the west flank of the anticline and the same two drag folds illustrated in the previous view.

Fan folds.—Folds of this type have been recognized at several places in the region. Usually they are so deeply eroded that only their stumps remain, or they are broken by faults. In the vicinity of Sugarloaf Mountain, however, in the Cranes Flat quadrangle, see Figure 8, there is a fine example of an inverted fan fold. The rocks immediately involved belong to the Homer limestone member of the Wayan formation (Lower Cretaceous?).

Sugarloaf Mountain was selected by St. John[1] of the Hayden Survey years ago as a station, and he drew a geologic structure section through it, in which he shows a southwesterly dipping series of strata, overlapped on the west by basalt at Sheep Mountain just west of the area shown in the figure, and arched into a prominent anticline at Sugarloaf Mountain by an igneous intrusion. The structure of the area near Sugarloaf Mountain is not so simple as figured by St. John. The limestone, which he did not differentiate from the other strata, is there thrown into a series of relatively sharp folds, among which narrow folds of the underlying sandstone rise to the level of erosion here and there. Southwest of Sugarloaf Mountain the dips of the limestone and the related strata are

[1] Orestes St. John, "Report of the Geological Field Work of the Teton Division," *U.S. Geol. and Geog. Survey Terr. (1877), 1879,* pp. 351–60.

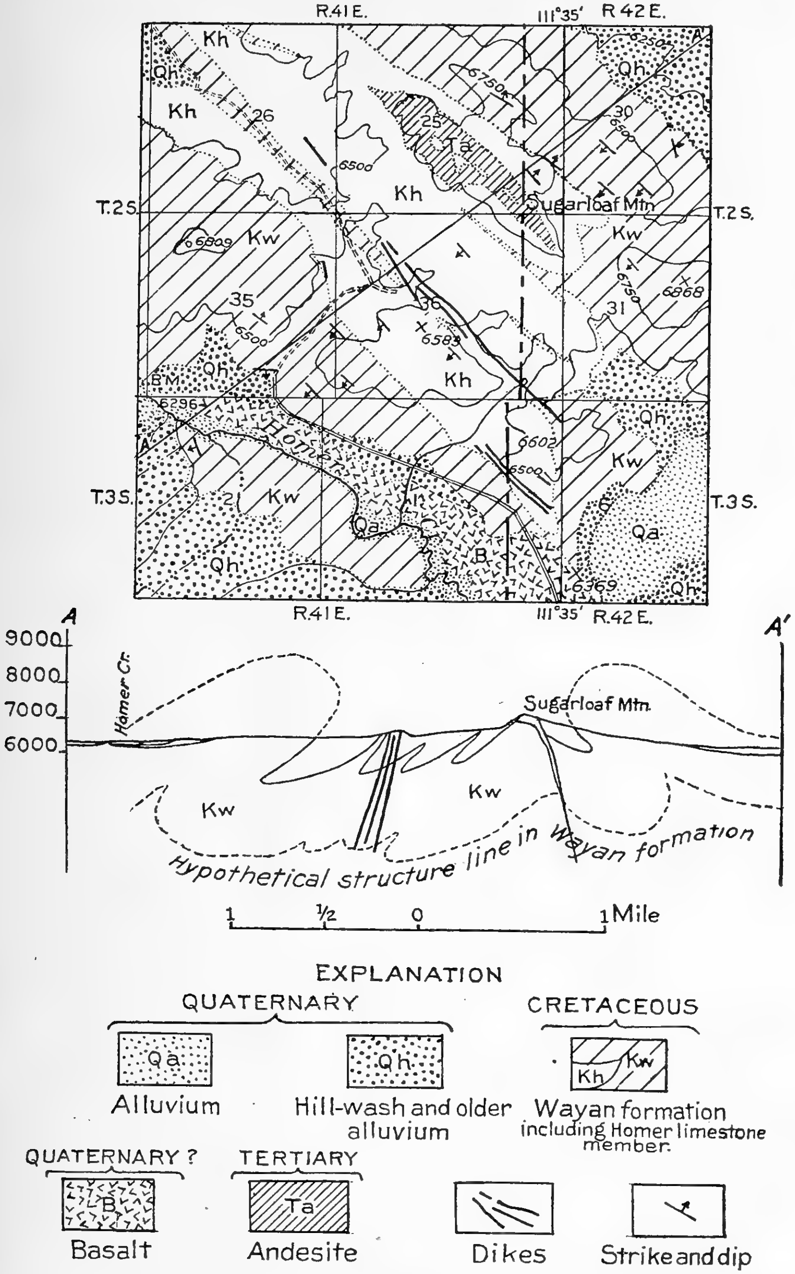


FIG. 8.—Map with geologic structure section of the Sugarloaf district, Cranes Flat quadrangle, showing an inverted fan fold.

southwest, but northeast of the mountain the dips of the limestone are northeast. Thus the structure of the limestone is a synclinorium, having the general form of an inverted fan fold. The intrusion at Sugarloaf Mountain is a thickened sill or incipient laccolith, arching with the strata in the northwestern extension of the mountain but eroded on the southwest limb beneath the summit and southeastward. The fold which forms the mountain is one of the subordinate folds of the synclinorium rather than a major structural feature supported by a relatively large intrusive body, as postulated by St. John. The geologic structure section *AA′* illustrates the features described above. Its line forms an angle of about 30° with that of St. John's section.

An example of what is believed to be the stump of an upright eroded fan fold is found in the western part of the Crow Creek quadrangle, see Figure 9. Snowdrift Mountain, part of one of the most persistent anticlines of the region, is flanked on either side by synclines which are in general inclined eastward. The Webster syncline on the east is markedly unsymmetrical, the west limb being steep and locally overturned eastward, but the east limb has a gentle westerly dip. The Georgetown syncline along the west side of Snowdrift Mountain is deeper and the limbs are steeper. The east limb is locally vertical or even overturned. Thus the intervening Snowdrift anticline is with little doubt an eroded fan fold. The structure sections along the lines *SS′* and *TT′* illustrate the features cited. At the line *SS′* the axis of the fan fold is somewhat inclined eastward and the Webster syncline is broken by a local thrust. Although the Snowdrift anticline does not everywhere show a tendency toward fan folding it is closely folded throughout most of its length and here and there exhibits that tendency, as shown on the west flank of Pelican Ridge, see structure section *GG′*, Figure 10*B*. The Dry Valley anticline, west of the Snowdrift anticline, see section *SS′* Figure 9*B*, locally has similar tendencies. Other instances which may not be figured here are illustrated in the forthcoming detailed report.

The Meadow Creek graben.—Perhaps the most striking effect of normal faulting in the region is the production of horst and graben structure in the northwestern part, see Figure 1c. The

valley of Meadow Creek in the southern part of the Cranes Flat quadrangle is a fault trough or graben. This structure may be traced about 15 miles southeast into the Lanes Creek quadrangle, where it apparently dies out. The northward extension of the graben is concealed by basalt and Quaternary deposits. The bounding ridges are composed of Carboniferous rocks and are conspicuous topographic features.

Two transverse normal faults intersect the graben, one near the south boundary of the Cranes Flat quadrangle and the other in the northwest corner of the Lanes Creek quadrangle, down-faulting the portion between them. The bounding ridges in the down-faulted area are farther apart than in the portions to the northwest or southeast. In the southeastern part of the graben beds of the Thaynes group (Lower Triassic) are exposed and in the widened, down-faulted portion both the Woodside shale (Lower Triassic) and Thaynes appear, though most of the area is underlain by basalt and Quaternary deposits. The structure of the rocks within the graben is probably synclinal, as shown in structure sections EE' and GG'. It is with little doubt the continuation of synclinal structures observed farther southeast.

The fault which lies along the northeast side of the graben is concealed for much of its length, but in the southeastern part of the area here shown is represented by two faults, separated by a narrow strip of the Phosphoria formation, but together bringing Lower Thaynes into proximity with the Wells. Northwest of the area illustrated the fault doubtless continues for some distance beneath the basalt. Its stratigraphic throw is not known but is estimated at 3,000 to 4,000 feet.

The transverse normal fault that passes between Limerock Mountain and Pelican Ridge causes the mountain to stand nearly a mile northeast of the line of continuation of the ridge. Similar effects in reverse order are produced where the fault intersects Little Gray Ridge. Neither the amount of the downthrow nor the hade of the transverse fault is known, but from the effects and assumed values for the dips of the lateral faults that bound the graben it is thought that 5,200 feet may represent the order of magnitude of the vertical displacement.

The other transverse normal fault intersects the graben southeast of Pelican Ridge. Its generally east-northeasterly course is largely concealed by basalt, but it cuts across Little Gray Ridge

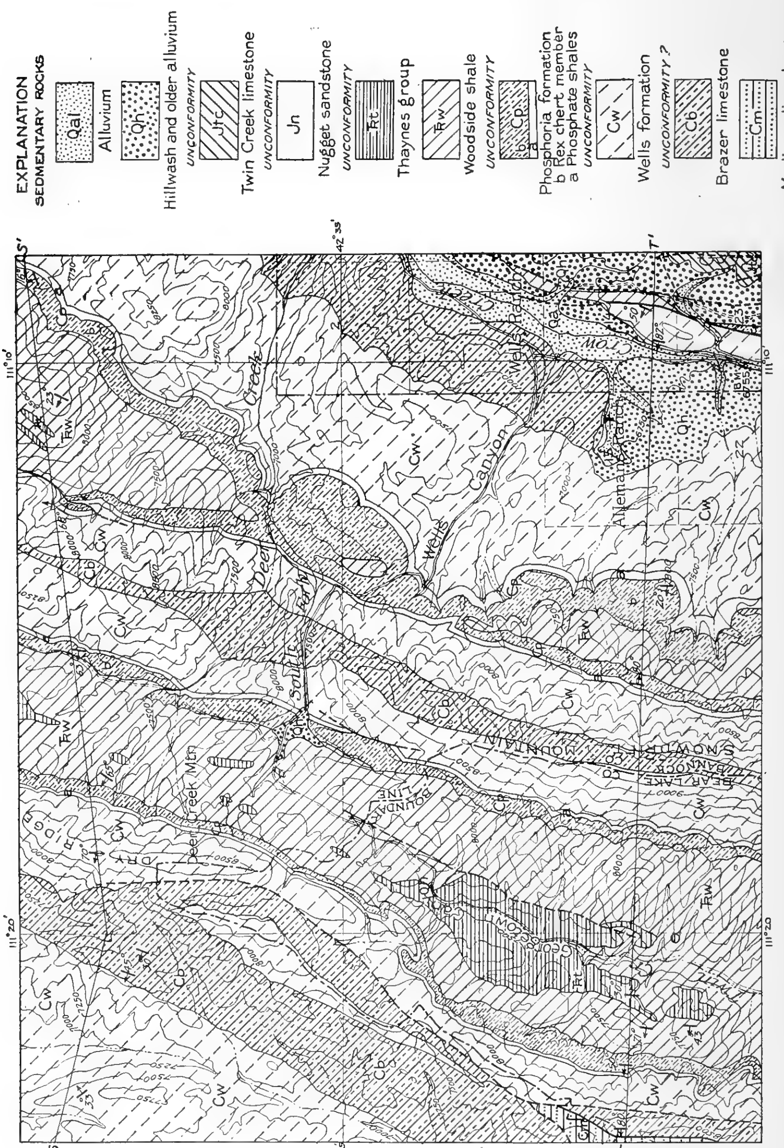


FIG. 9A.—Map of parts of the Preuss Range, Slug Creek, and Crow Creek quadrangles, showing eroded fan folds and associated features.

and there offsets the boundary between the Brazer and Madison limestones (Mississippian). Basalt has outflowed on the west flank of the ridge along part of the fault trace. The downthrow

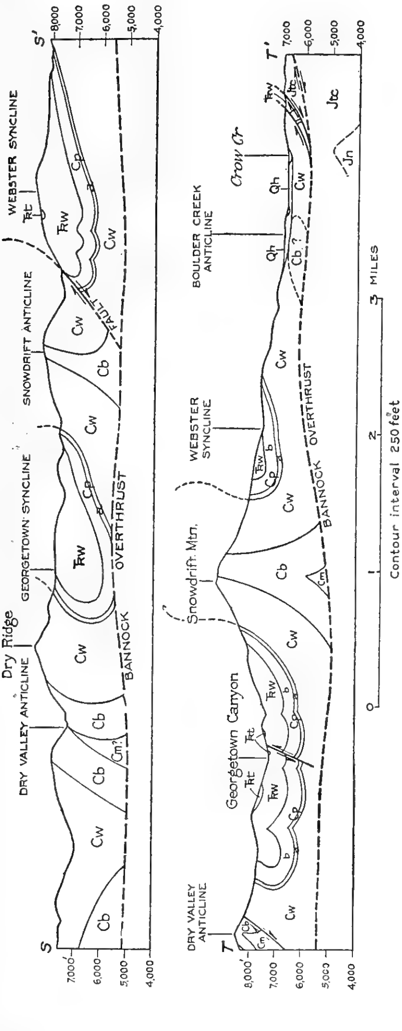


FIG. 9 *B*.—Sections along the lines *S–S'* and *T–T'* of Fig. 9 *A*

is to the northwest and, employing similar assumptions to those noted for the last described fault, an estimate of 3,500 feet may be made for the vertical displacement at this locality.

The faulted ridge including Limerock Mountain and Pelican Ridge, a continuation of the Snowdrift anticline mentioned above, is here a horst, for it stands between the Meadow Creek graben on

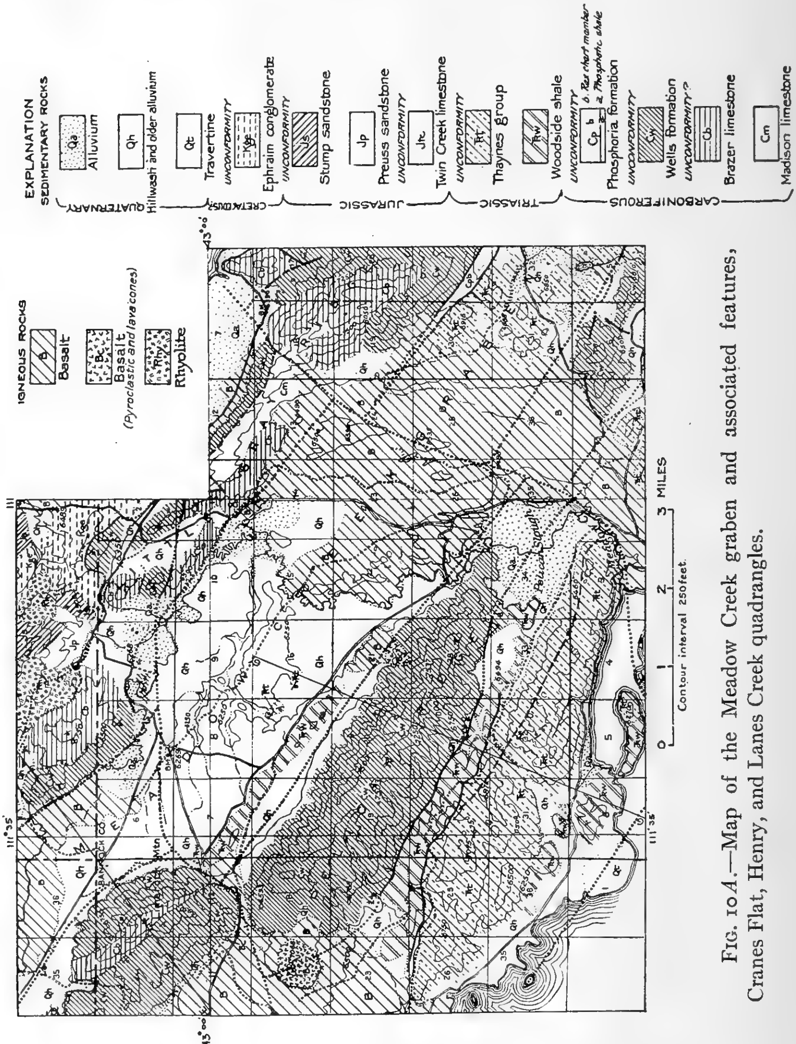


FIG. 10A.—Map of the Meadow Creek graben and associated features, Cranes Flat, Henry, and Lanes Creek quadrangles.

the northeast and another down-faulted area on the southwest. The relations of this horst to the adjoining areas, both northeast and southwest, are shown in structure sections *EE′* and *GG′*.

The ridge northeast of the graben is not clearly a horst though it is much broken by faults and has suffered extrusion of rhyolite.

NOTES ON THE DEFORMATION OF SOUTHEASTERN IDAHO

Epochs of deformation.—Although southeastern Idaho was profoundly affected by crustal disturbances at the close of the Jurassic, the observed mountain structures appear to be the result of two later epochs of mountain building. The earlier of these occurred after the deposition of the Wayan formation and before the deposition of the Wasatch formation. It probably corresponds

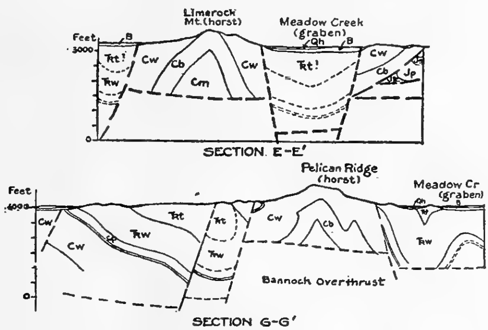


FIG. 10*B*.—Sections along the lines *E–E'* and *G–G'* of Fig. 10*A*

with the interval between the Adaville and Evanston formations of Veatch[1] or the epoch which, according to Ransome,[2] "appears to have begun at the close of the recognized Laramie or possibly even earlier, and to have attained its maximum between the Fort Union, which chiefly on the basis of its plant remains is generally classed as basal Eocene, and the mammal-bearing lower Eocene Wasatch."

The second mountain-building epoch occurred after the deposition of the Salt Lake formation which, on the basis of present rather unsatisfactory evidence, is tentatively assigned to the Pliocene. This formation locally has steep dips thought to have been produced by deformation in late Pliocene or post-Pliocene time.

[1] A. C. Veatch, "Geography and Geology of a Portion of Southwestern Wyoming," *U.S. Geol. Survey Prof. Paper 56* (1907), p. 75.

[2] F. L. Ransome, "The Tertiary Orogeny of the North American Cordillera and Its Problems," *Problems of American Geology*, pp. 287–376, p. 322, New Haven, 1915.

Rocky Mountain geosyncline.—Southeastern Idaho forms a part of a great geosyncline in which sediments were deposited with few interruptions of magnitude from early Cambrian to Upper Cretaceous times. This great trough extended from the Arctic Ocean southward through the Great Basin and was in general an area of subsidence or a negative element[1] on which the sediments had accumulated in great thickness. On the west during the same interval a relatively persistent land mass or positive element had separated the geosyncline from the Pacific Ocean, and on the east a less persistent barrier at times had separated it from the interior sea.

The geosyncline served to localize the deformation and had a directive influence upon it. The tangential pressure which produced the folds and overthrusts was normal to this structure and, in southeastern Idaho, came from the west southwest.

Initial dips within the geosyncline and differences in the character of the sediments doubtless tended still further to localize the folds and thrusts and to determine their character.

Favorable formations.—Many of the Paleozoic formations are massively bedded and would act as competent strata under deformation. A number of formations, however, contain shaly members. Some of the limestones, too, are thin bedded. Such formations exposed to deformation in the zone of fracture would furnish horizons in which thrust planes might originate. The Bannock overthrust zone is complex and no one formation has yet been identified as the source of the thrust plane.

The Mesozoic formations are generally weaker and less well consolidated than are the Paleozoic rocks. Lying with favorable initial dip and in great thickness athwart the direction of maximum compression, the Mesozoic rocks crumpled under the accumulating compressive stress and permitted the more or less folded Paleozoic rocks with some accompanying or overlying Mesozoic rocks to override them. They thus generally form the basement over which the great thrust block of the Bannock overthrust moved and on which it now rests. Although it has been customary in

[1] Bailey Willis, "A Theory of Continental Structure Applied to North America," *Bull. Geol. Soc. America*, Vol. XVIII (1907), pp. 389–412.

the discussion of overthrusts to regard the lower block as passively overridden by the upper or thrust block, it is probable that both participate in the movement, the separated parts moving past each other, as suggested by Barrell.[1]

Horizontal thrusting.—The original nearly horizontal attitude of the Bannock thrust plane has been modified by subsequent compression and folding, but it indicates that the effective deformative forces acted horizontally and were not the surface expression of obliquely emerging, deep-seated shear, such as was postulated by Willis[2] for the fault zone along the east side of the Sierra Nevada Mountains.

Factors in deformation.—Chamberlin and Miller[3] have shown from their own experiments and from the earlier work of Cadell, Willis, Adams, and others that many factors are involved in the production of low-angle faulting, such as is exemplified in great overthrusts. Among these may be mentioned: (1) rotational strain; (2) increase in resistance to deformation with depth; and (3) a relatively large ratio of thrust to weight.

(1) Rotational strain as a factor in the deformation of southeastern Idaho is clearly indicated by the frequency of inclined or overturned structures.

(2) Although no data are available regarding conditions in depth, it is clear from the horizontality of the thrusting previously mentioned and from the locally fractured and generally unmetamorphosed condition of the strata, that the deformation took place at no great depth. The visible structures at least were developed in the zone of fracture. No evidence of flowage has been found.

(3) The great horizontal displacement produced by the Bannock overthrust shows that the thrust was enormous. The weight, on the other hand, could not have been very great because of the apparent shallowness of the deformation.

[1] Joseph Barrell, "The Upper Devonian Delta of the Appalachian Geosyncline," *Am. Jour. Sci.* (4th ser.), Vol. XXXVII (1914), p. 107.

[2] Bailey Willis, "Structure of the Pacific Ranges, California," *Bull. Geol. Soc. America*, Vol. XXX (1919), pp. 84–86.

[3] R. T. Chamberlin and W. Z. Miller, "Low-Angle Faulting," *Jour. Geol.*, Vol. XXVI, No. 1 (1918), pp. 1–44.

Later deformative epoch.—The later epoch of deformation was marked by broad uplift rather than by intensive folding. There was, however, some folding, involving locally steep or overturned dips, but generally of an open character. The folding of the plane of the Bannock overthrust is of this type and is thus structurally more akin to the later than to the earlier deformative epoch.

Relaxation and readjustment.—The vigorous compression of the earlier deformative epoch was succeeded by relaxational phases involving normal faulting and gradual readjustment to new conditions of equilibrium. Some normal faults along the Bannock fault zone represent with little doubt the fracture and jostling of blocks under light load near the margin of the fault block. Other normal faults partly concealed by Tertiary beds also may be referred to the interval of relaxation following the earlier deformative epoch. Many of the normal faults, however, including those that produced the horst and graben structure, are not associated with Tertiary beds. There is a single doubtful exception to this statement. On the other hand there is definite evidence that Tertiary beds have been displaced by normal faults. Thus these faults have been considered as later than the Tertiary beds but earlier than the earliest Quaternary, and hence associated with the relaxational interval succeeding the later deformational epoch.

The horst and graben structure is more or less intimately associated with extrusions of ryholite and basalt. The basalt in particular has flooded the valleys in the vicinity of these structures and has emerged along some of the fault lines. On the other hand, sufficient erosion had occurred after the faulting to produce practically the present topography before the extrusion of the basalt. That event, therefore, probably accompanied a relatively late reopening of some of these faults, together with the development of new fissures.

At present no definite evaluation of the parts to be assigned to the two relaxational intervals may be made.

DISCUSSION OF "SUMMARIES OF PRE-CAMBRIAN LITERATURE OF NORTH AMERICA," BY EDWARD STEIDTMANN

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The recent "Summaries of Pre-Cambrian Literature of North America" by Professor Edward Steidtmann[1] bring up points to which certain contributions and corrections should be made. He states that,

> (1) . . . . the pre-Cambrian rocks northeast of Lake Huron show one conspicuous unconformity, and above the conspicuous unconformity are two series of slightly metamorphosed, dominantly clastic sediments, separated by an inconspicuous unconformity. (2) The lower one, the Bruce series, locally contains tillites. The upper series is generally known as the Cobalt series. (3) At Killarney, on the north shore of Lake Huron, Collins has found that the Bruce and possibly the Cobalt series are intruded by Killarney granite, and in this locality they assume many of the characteristics of the older series, the Timiskaming.[2]

In regard to the so-called "inconspicuous unconformity," Steidtmann followed Coleman (1915)[3] and Miller and Knight (1915).[4] Later work by Collins[5] shows that in the region southeast of McCabe Lake, north of Cutler, erosion removed the Serpent quartzite, the Española group, the Bruce conglomerate, and part of the Mississagi formation, amounting to the greater part of the Bruce series, previous to deposition of the Cobalt series. In 1914 he showed that the erosion interval amounted to thousands of

[1] *Jour. Geol.*, Vol. XXVIII (1920), pp. 643–58.

[2] *Ibid.*, p. 643. The numbers have been inserted by the writer.

[3] A. P. Coleman, *Problems of American Geology*, pp. 81–161. Quoted by Edward Steidtmann, *op. cit.*, p. 647.

[4] W. G. Miller and C. W. Knight, *Jour. Geol.*, Vol. XXIII (1915), pp. 585–99. Quoted by Edward Steidtmann, *op. cit.*, p. 656.

[5] W. H. Collins, report in preparation.

feet over large areas.[1] And in the Española[2] area there is a probable difference of erosion of more than 5,000 feet within 10 miles. It is largely a matter of opinion as to how important such an unconformity should be considered, but it is surely not inconspicuous. Erosion intervals of thousands of feet in Paleozoic series are considered notable, and they should not be undervalued in pre-Cambrian series. However, in fairness to all it must be added that there are places where the unconformity appears to be slight; generally only by tracing it over large areas, as Collins has done, can the observer recognize its true greatness. Furthermore, in comparison with the great unconformities beneath and above the Huronian formations the Bruce-Cobalt unconformity is much less conspicuous. This at least may be said, the unconformity locally is inconspicuous but nevertheless important.

Second, in saying that the Bruce series, the lower Huronian series of Ontario, locally contains tillites, Steidtmann appears to follow Coleman (1915).[3] The fact is that the tillites are characteristic only of the upper group, the Cobalt series, or specifically the Gowganda formation of Collins.[4] The glacial origin of at least part of the Cobalt conglomerate is held by Wilson (1913),[5] Collins (1914),[6] and by Coleman himself first (1907)[7] and last (1920).[8] The Bruce conglomerates in certain phases are distinctly different in character from the Cobalt tillite. They lack the thinly laminated slates and the well-bedded slate layers carrying scattered

[1] W. H. Collins, *Canada Geol. Survey, Museum Bull. No. 8* (1914), p. 21.

[2] T. T. Quirke, *Canada Geol. Survey, Mem. No. 102* (1917), p. 42. Apparently through errors in copying, the summary of Quirke's classification on p. 657 of Steidtmann's paper differs considerably from the work summarized, both by omissions and by faulty arrangement. See *Canada Geol. Survey, Mem. No. 102*, pp. 6 and 7.

[3] Quoted by Steidtmann, *op. cit.*, pp. 646, 648.

[4] W. H. Collins, *Canada Geol. Survey, Mem. No. 95* (1917), p. 10.

[5] Morley E. Wilson, *Jour. Geol.*, Vol. XXI (1913), pp. 121–41, and *Canada Geol. Survey, Mem. No. 17* (1912). Quoted by Steidtmann, *op. cit.*, p. 658.

[6] W. H. Collins, *Congrès géologique international*, XIIth Session (1914), pp. 399–407. Quoted by Steidtmann, *op. cit.*, p. 650.

[7] A. P. Coleman, *Am. Jour. Sci.*, Vol. XXIII (1907), pp. 187–92; *Jour. Geol.*, Vol. XVI (1908), pp. 149–58; *Bull. Geol. Soc. Am.*, Vol. XIX (1908), pp. 347–66.

[8] A. P. Coleman, *Economic Geology*, Vol. XV (1920), No. 6, pp. 539–41.

pebbles and bowlders. They have little argillaceous matrix about the inclusions in massive conglomerates. They are not "slate" conglomerates, as part of the Gowganda formation is. Their character is essentially that of a basal or alluvial deposit. In part they are well sorted and in part they are massive, but the massive material is composed characteristically of dark-colored graywacke, or of gritty or bowldery conglomerate. In no instance is there record of striated, glacially soled bowlders being found in Bruce conglomerates. Nowhere to our present knowledge is there a polished basement beneath them. The matrix of the Bruce conglomerates is commonly dark-colored, arkosic, and graywacke-like; the matrix of typical Cobalt tillite is green, pale-green on weathered surfaces, and looks like metamorphosed clay. So clear is the difference in the character of certain phases of the matrices that members of Collins' parties from 1914 to 1918, with some practice, were able to tell from a glance at typical hand specimens whether or not the rock was Bruce or Cobalt conglomerate. This difference in the character of the matrix is a genetic difference, connected with the glacial origin of the one and the non-glacial origin of the other. Those phases of the Bruce conglomerate which are very similar to the less characteristic phases of the Cobalt conglomerate conceivably may be of an obscure, glacial origin.

It is not altogether surprising that a reviewer of the literature should fall into confusion. Originally, in accordance with the best nomenclature of the day, Coleman[1] referred to the Cobalt conglomerate as Lower Huronian, as is quite clear from his writings in 1907 and 1908; whereas the only locality of distinctly glacial deposits he cited is the Cobalt silver-producing district which is underlain by the upper, Cobalt series, not by the Bruce conglomerate now known as the lower series. Furthermore, Coleman, himself, at one time seems to have been confused by the similarities between the Bruce and the Cobalt conglomerates. Indeed Professor Willmott[2] previously (1901) had written: "The two slate

[1] A. P. Coleman, "The Lower Huronian Ice Age," *Jour. Geol.*, Vol. XVI (1908). p. 149; *Bull. Geol. Soc. Am.*, Vol. XIX (1908), p. 355; *Am. Jour. Sci.*, Ser. 4, Vol. XXIII (1907), pp. 190–91.

[2] A. B. Willmott, *American Geologist*, Vol. XXVIII (1901), p. 19.

conglomerates of Murray are so much alike that they cannot be distinguished. Where the limestone band is absent, as it often is, they join, and Murray himself confesses that he could not draw the dividing line." Nevertheless, in some such places the dividing line may be, and has been, drawn. On the other hand, there are phases of the Bruce conglomerate so similar in character to phases of the Cobalt conglomerate that no distinctions have yet been recognized. Another factor which may have caused confusion in the reports of Coleman is the fact that he found at Cobalt the Cobalt conglomerate to be the basal conglomerate of the Huronian formations, the entire Bruce series being wanting. Thus, carrying his correlations westward from Cobalt, he supposed the basal conglomerate of the original Huronian area to be the same as that at Cobalt, whereas it is actually the base of the Mississagi formation of the Bruce series. However, all this was clearly put straight by Collins[1] in 1916, and it seems a pity to have confusion again after the known facts have been published. So far as is now known, the Bruce conglomerates, certainly for the main part, are not of glacial origin, but some of the Cobalt conglomerates are agreed to be tillites.

Regarding the last topic, the work of the writer carried on this summer near Lake Geneva, 20 miles northwest of Sudbury, Ontario, shows that syenitic masses intrude the Cobalt series, thus confirming and complementing the work of Collins (1916)[2] on the age of the Killarney granite. Collins found that the Bruce series certainly and possibly the Cobalt formations, are intruded by an acid intrusive in an area north of Lake Huron, from 15 to 25 miles southward from Sudbury. Now it is known that what might have been considered a local phenomenon of little consequence in pre-Cambrian classification and correlation must be regarded as probably a widespread and considerable intrusion. The age of these intrusions having been determined and confirmed in areas 40 miles apart, it becomes necessary to scrutinize carefully those local correlations and distinctions which are based largely upon different periods of orogenic movement and acid intrusions. Almost certainly it will

[1] W. H. Collins, *Canada Geol. Survey, Mus. Bull. No. 8* (1916).

[2] *Ibid. No. 22* (Feb. 5, 1916). Quoted by Steidtmann, *op. cit.*, p. 650.

be recognized, as it has been found already, that some masses of supposedly pre-Huronian and Sudburian rocks will be identified as Huronian sediments intruded by these late pre-Cambrian granites and syenites. However, there are surely some areas which are not subject to such a revision, in which the reality of the pre-Huronian sediments seems to be beyond dispute; so that we may bring our ideas of the pre-Cambrian succession of events north of Georgian Bay more up to date, as follows (the events being listed in chronological order from the bottom up):

SEQUENCE OF PRE-CAMBRIAN EVENTS IN THE TIMISKAMING REGION

Proterozoic Era
- Intrusions of Killarney and Geneva granites and syenites, accompanied by severe faulting, mountain folding, and extensive warping.
- Injections and extrusions of basic rocks (Keweenawan)
- Deposition of Whitewater series
  - Chelmsford sandstone
  - Onwatin slate
  - Onaping tuff
  - Trout Lake conglomerate (not tillite)
  - Hiatus, relations unknown
- Deposition of Cobalt series
  - White quartzite
  - Cherty quartzite
  - Lorrain quartzite and conglomerate (not tillite)
  - Gowganda formation—including tillites
  - Considerable interval of erosion, the resulting sediments being unknown
- Deposition of Bruce series
  - Serpent quartzite and conglomerate (not tillite)
  - Española limestone
  - Española graywacke
  - Bruce limestone
  - Bruce conglomerate (probably not tillite)
  - Mississagi quartzite and basal conglomerate (not tillite)
  - Great interval of erosion, the resulting sediments being unknown

Archeozoic Eras
- Time of orogenic diastrophism accompanied by acid intrusions (Algoman)
- Deposition of pre-Huronian sediments (Sudburian and others), quartzites, graywackes, conglomerates

Great interval of erosion, the resulting sediments being generally unknown, but represented in part by Sudburian and other pre-Huronian sediments
Time of granite intrusions and diastrophism (Laurentian—all inferred from the presence of granite bowlders in pre-Huronian conglomerates)
Deposition of products of Keewatin weathering, accompanied and interrupted by volcanic extrusions and intrusions of undetermined order and distribution

NOTE:—The above communication has been made with the permission of the Director of the Geological Survey of Canada.

THE NATURE OF A SPECIES IN PALEONTOLOGY, AND A NEW KIND OF TYPE SPECIMEN

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In discussing the subject, "What is a species?" with my associates in vertebrate paleontology, many points of interest have come up which will bear repeating. The great complexity of the problem and the number of elements which enter into its composition make the solution difficult—perhaps impossible—for certainly no single definition can apply in every case.

Professor Schuchert has said that he finds it best to let the man who is studying a given specimen decide what a species is, after he has read several definitions, and that even for students who are writing their first papers it is best left to the individual; he may stand or fall on his ingenuity in handling this unanswerable question.

The more deeply we go into the study of a specimen, not only do the features multiply, but they increase in importance, and thus we raise a variety to a species, or a species to a genus. Along with the personal element and the extent of our knowledge, very important factors indeed, we must in paleontology consider another, viz., the actual worth of the specimen, and it is undeniable that a new species should not be made on fragmentary material unless its characters are unique and of such importance as to reveal the existence of a new and strange group of animals, hence scarcely less than a genus or even a family. As an example of such specimens, whose discovery gave us the first knowledge of strange groups, there may be mentioned *Archaeotherium mortoni* Leidy, *Ammodon leidyanus* Marsh, or *Rhinoceros occidentalis* Leidy.

It is a well-known fact, and yet one not fully recognized, that new types based on good specimens are and should be made, to supplant older inadequate ones. We may cite as an example the mammal *Dinohyus* Peterson, which Matthew says is not different

from *Daeodon* (*D. shoshonensis* Cope), or the dinosaur *Tyrannosaurus*, which may be identical with *Manospondylus* (*M. gigas* Cope). Many such examples might be given showing where a more complete skeleton has usurped the taxonomic position of a type based on a mere fragment, thereby forcing the latter to fall into a group of historic relics, which merely mark the progress of our science.

A new kind of type specimen.—It is here suggested that we recognize this principle of substitution as a useful one, and we propose the name of *protype* (=for the proterotype) for such a supplanting type specimen. A protype should be based only on a very complete skull or skeleton. It is based on supplementary material, and as such is one of the forms of apotypes. In other words, a protype may be said to be a proxy, for it operates with full authority under a special designation. The material of a protype should, it would seem, be given a new name, preferably as a subspecies, for in such a humble position under the original proterotype it preserves the name of the older species replaced; it has the taxonomic advantage of linking the two together, and furthermore, if it is later found that the new specimen is not identical with the older, the subspecies can be raised to the full rank of a species.

Thus it is evident that a new species name should only be given, first, in the case of a protype, as just set forth, and second, to any specimen, fragmentary or perfect, which clearly possesses distinguishing features or some unusual or unique morphologic character that may be accentuated by its stratigraphic occurrence and that needs to be published and made widely known, thereby adding to the sum total of human knowledge.

Making of species.—Many naturalists have ventured an opinion as to what limit of variation should constitute the bounds of a species. This has, by common consent, been determined by the limits of interbreeding, where procreation is impossible or the offspring is sterile. In the natural state, it is interesting to note that, though the barrier is usually a physical or physiological one, yet it may be purely psychological, and as such may break down in captivity, giving rise to most unusual hybrids capable of reproducing their kind.

There exist the so-called geographical species which show no marked morphologic differences. Modern biologists make such distinctions in color or in the nature of the hair or muscles; these are, however, criteria necessarily lost to the paleontologist, who generally has only the hard parts for study. The student of vertebrate fossils is further limited, as a rule, in his making of species, to a single specimen, while the student of modern biology has many cotypes or paratypes.

Considering the vastness of geological time, it is easy to see that two specimens of the same geological formation may have been separated by tens of thousands of years, during which important changes not only of habitat but as well of form and habits may have occurred. In such a case, nothing of the resulting trivial changes in the soft parts may be discerned.

Relativity of species.—Generally the small unit characters have value only when grouped; not always does a specific feature fully determine the bounds of a species. But on the other hand, no matter how trivial this character may appear to be, if we find it occurring constantly in one set of specimens and not in others, it is to be regarded as specific in value and typifies the individuals of that group. Furthermore, not all the varieties referred to a species may have all of the so-called specific characters. Two specimens under the same species may have but one or two only of the several secondary features supposed to characterize the group as a whole.

Distinguishing characters are of two classes, relative and absolute. Sometimes a relative difference may be advanced so far that it appears to be, or may actually become, absolute; for example, the growth of a bone until it meets a joint and develops a facet, or the reduction of a premolar to a point where it becomes obsolete. The so-called absolute characters may be thought of as unit characters. Relative characters may be illustrated by the familiar variation in size, and then we wonder what the limits of size are beyond which animals cannot interbreed, for we realize how positive a barrier this may become. In this respect it seems safe to say that in mammals a size difference of 30 per cent would

tend to separate two groups, based on our general conception of the ultimate barrier between species.

Carrying the subject to a further quantitative analysis, what range of "ratios" can we allow within a species? Two skulls of approximately the same size may vary in proportions, one part being 15 per cent larger, another part as much smaller. Here is a range of 30 per cent; is it a specific difference? The personal equation of course enters into all such questions and the resulting taxonomy.

Presenting specific characters.—A glance through the literature shows many instances where an author at the beginning of an original description states that the "specific characters" are so and so, and proceeds to list a number of features which, taken singly, might not even be subspecific, or, on the other hand, might be generic. In addition, these "specific characters" which may serve to distinguish two species do not show a contrast with a third or fourth; for instance, characters "a" may show the distinction from species "A," but we must look for another group of criteria, "b," to separate our species from "B." A species depends not only upon its own features as selected by its author, but upon the features of the other species in the genus as well; the specific characters marking the boundary in one direction may not be such as to show it in another.

New discoveries are constantly being made which contradict general statements of distinctions, and overturn our nicely adjusted taxonomy, unless we carefully indicate the type material compared when we draw our contrasts. The safer course, then, seems to be to limit one's self to the "distinguishing characters," as so many paleontologists do, without making a guess as to what the undiscovered or unknown specimens may show; or to indicate just what species are being distinguished by certain characters, in which case the description resolves itself into as many parts as there are species to be compared, and each description shows definite contrasts and possibly a wholly different set of distinguishing features; all of this is again subject to one's learning and powers of contrasting the variables.

Summary.—In the last analysis it rests with each author what his specific differentiations will be, and their validity will often depend upon and be rated by the general standard of his work.

It is proposed here that, in accordance with a recognized need, we employ *protypes* to serve as "proxy types" to substitute for inadequate existing type material.

The practice is criticized of attempting to classify the characters of a type specimen as specific or generic without knowing all the related forms or without specifying the relation to each neighboring species or genus. It is suggested that contrasts be drawn with each other species separately, or that the noncommittal term "distinctive features" be used.

Reviews

Het Verband tusschen den plistoceenen Ijstijd en het Ontstaan der Soenda-Zee (Java- en Zuid-Chineesche Zee) en de Invloed daarvan op de Verspreiding der Koraalriffen. . . . . (*The Sunda Sea and Its Barrier Reef.*) Door G. A. F. Molengraaff. . . . . K. Akad. Wet. Amsterdam, Verslag der Afdeeling Natuurk., Vol. XXVIII, 1919, pp. 497–533.

The shallow Java Sea between Java-Sumatra and Borneo and the confluent shallow southern part of the China Sea are united by Molengraaff under a single name, the Sunda Sea. He accounts for the flat sea floor, nowhere more than 40 fathoms deep, by supposing that, so far as its area was already a lowland of weak rocks in pre-Glacial time, it was worn down still lower during the Glacial epochs of lowered sea surface, and that its deeper, sea-covered parts were in the same epochs filled up to a corresponding level. Large rivers, fed by the heavy rainfall of the region, are thought to have been active agents of gradation. The degraded and aggraded lowland was submerged and the present sea created when the ocean finally rose to its normal level in post-Glacial time. The margin of the submerged lowland is assumed to lie at the present 40-fathom line, outside of which a relatively rapid descent is made to deep water. Evidence of submergence is found not only in the embayed mouths of tributary rivers not yet filled with deltas, but also in the occurrence of detrital tin ore in the extension of river courses a mile or more from the shore of certain tin-bearing islands. The fine sediments by which the sea bottom is now covered are regarded as river deposits laid down while the sea was rising to its present level.

Although the region here considered—Sundaland, as Molengraaff calls it—is known on the basis of abundant geological evidence to have long been much more stable than the disturbed region of the several deep seas and many islands farther east, the essential exclusion of crustal subsidence and the restriction of the Sunda sea-floor origin to so short a geological interval as the Glacial epochs of the Glacial period seem open to question. Moreover, the present margin of the shallow floor near the 40-fathom line, which Molengraaff regards as the built-out border of the worn-down lowland when the Glacial ocean was lowered,

is open to another interpretation. It may be explained as the outer margin of a continental shelf that has been gradually forming through late Tertiary and Quaternary time, the final touches having been given to it in post-Glacial time by the action of waves and currents on the fine silts that are so abundantly supplied by the inflowing rivers; the sea water is today sometimes discolored 60 km. from the river mouths. It may be recalled in this connection that Daly has found that "the charts of the world show the break of slope on the [continental] shelves to be near the 40-fathom line," and that it is only to this depth that "waves and currents are competent to advance the outer edge of a continental embankment with noteworthy speed." Hence, instead of interpreting the 40-fathom edge of the Sunda Sea bottom and the edges of many other continental shelves of like depth as marking continental shore lines when the ocean was lowered in the Glacial epochs—the ocean lowering being calculated, singularly enough, to have been but little less than the same measure of 40 fathoms—these shelf edges are better interpreted as being due in significant measure to the adjustment of the submarine profile to normal sea-level by submarine processes in post-Glacial time.

Coral reefs are rare or wanting in the Sunda Sea by reason of the muddiness of its bottom and the resulting impurity of its waters, as Sluiter explained thirty years ago. But the margin of the eastern part of the sea floor, the so-called Borneo bank, is occupied by an imperfect and discontinuous barrier reef for over 450 km. Molengraaff points out that the reef is separated at its easternmost extension only by the 50- or 60-km. width of the deep-water Macassar Strait from the island of Celebes, while it is separated by 160 km. of shallow water from Borneo; hence it is related not to the nearer but to the farther one of these islands. The imperfection of the reef is ascribed to the muddiness of the sea-floor margin on which it is assumed to have originated as the ocean rose in post-Glacial time; but the muddiness of the shallow sea floor is so great that it is doubtful whether any reef could have originated there at all. The reef may perhaps be better explained as a recent upgrowth from such parts of a Tertiary reef as were not overwhelmed and destroyed by detrital deposits in the extension of the shallow sea floor while the ocean was lowered in Quaternary time.

Mention may be made of a number of imperfect atoll reefs which Molengraaff describes as bordering several extensive 20–40-fathom shoals that rise from deep water east of the Sunda Sea, in the region between the stable western islands and the unstable eastern ones. The

chief reefs here are on the Kalu-Kalukuang bank, 98 by 58 km.; on the several Laars banks, about 65 km. in total length; on the Postillion bank, 140 by 50 km.; and on the Paternoster bank, 115 by 26 km. The 40-fathom depth of the bank margins is ascribed by Molengraaff, following Daly's Glacial-control theory of coral reefs, to abrasion during the Glacial epochs of lowered sea-level, particularly to the phases of rising sea-level; but it may be more plausibly ascribed to wave-and-current aggradation with reference to normal sea-level in post-Glacial time, as above suggested. For even if a bank 50 or more km. in diameter were cut away by the waves during the lowered stand of the ocean in the Glacial period, its margin ought to have been at least 20 or 30 fathoms below the sea-level of that time and hence not 40 but 55 or 65 fathoms below present sea-level, with a gradual shoaling toward its center. Departures from such a form should therefore, under the explanation of the banks by abrasion, be accounted for by post-Glacial reef growth and submarine aggradation. In any case, it would appear that, if no disturbance takes place, these imperfectly reef-rimmed banks will in time develop into typical atolls of large size.

W. M. D.

Atollen in den Nederlandsch-Oost-Indischen Archipel. De Riffen in de Groep der Toekang Besi-Eilanden. (*Atolls in the Dutch East Indies.*) Door Dr. B. G. Escher. Batavia, Java: Mededeelingen . . . . Encycl. Bureau, Vol. XXII, 1920, pp. 7–18.

There has been discussion for some years past among the geologists of the Dutch East Indies as to the occurrence or absence of true atolls in their archipelago. Although atolls are certainly rare in that region, the occurrence of several typical examples is made clear by Escher, who had opportunity in March, 1919, of examining several reefs in the Tukang Besi group, southeast of Celebes. A copy of the original survey of the islands by the Hydrographic Service of the Dutch East Indies on a scale of 1:200,000, containing a greater number of soundings than those represented in the chart published for the use of navigators, is reproduced in Escher's paper. It shows seven atolls, the smallest about 2 km. in diameter; the largest measuring 48 by 15 km. and inclosing a lagoon 15 or 20 fathoms deep. Escher points out that these atolls lie in two belts, trending northwest-southeast, 100 or 130 km. in length, and that between the two belts and to the northeast of both of them are two roughly parallel belts of high islands bearing raised reefs, with fringing reefs at sea-level. The overall breadth of the four belts is about 140

km. These facts lead him to conclude that the two atoll belts have subsided, while the two high-island belts have risen; in a word, that the region has suffered a gentle folding, the atolls growing upward in the faint synclines. True-scale profiles show the exterior slopes of the atolls to vary from 30° to 69° down to depths of from 100 to 400 fathoms. It may be added that the prevailing absence of atolls in the deep seas inclosed by the islands of the East Indian archipelago is plausibly explained by the too rapid subsidence of the sea bottoms and of any islands that may have risen from their deeper parts in that very unstable part of the earth's crust.

W. M. D.

Les Iles Wallis et Horn. (*The Wallis and Horne Islands, Pacific Ocean.*) Par le Dr. M. Viala. Bull. Soc. Neuchât. de Géogr., Vol. XXVIII (1919), pp. 209–83. With halftone plates and an outline map of Wallis, 1:60,000.

The author of the above-cited article served as resident physician on the islands, which are French possessions, from 1905 to 1909; his geographical descriptions are general; his notes on the natives are much more detailed. Wallis, northeast of Fiji, consists chiefly of a main island, Uvea, of volcanic origin, 18 km. long by 6 or 8 km. wide, and about 200 m. in altitude; but there are also nineteen small satellite islands close by, of which three are volcanic, and the others are of coral origin. About half of the latter stand on the fine barrier reef, which, about 100 m. broad and interrupted by only four narrow passes, encircles the main island. The inclosed lagoon is from 2 to 5 km. wide, and is much interrupted by shoals: its depth is not stated. A well-formed fringing reef surrounds Uvea, so that canoes can reach the shore only at high tide. A wharf for larger vessels is built across the fringing reef at the chief village. While the low coral-sand islands are covered with luxuriant vegetation, the uplands of the main island have an infertile clayey soil and bear but scanty vegetation, chiefly ferns; except that a few cavities, interpreted as ancient craters and about 50 m. deep, have a richer growth; one such cavity contains a small lake. The uplands descend to an irregular shore line, where sand flats, often inclosing shallow lagoons of small size, afford the only cultivable ground; here the villages lie and here the coco-nut palm flourishes, yielding the most important commercial product of the islands; but rats abound and injure the crop. There are no streams, but springs emerge at the inner border of the sand flats. The southeast trade wind, blowing continuously and often

with violence from April to October, gives fair weather and leads the European residents to occupy the eastern coast of the island. From November to April the wind is light and variable with not infrequent calms; rains are then heavy and the high humidity makes the weather next to unbearable.

The Horne Islands are about 250 km. nearer Fiji; but as they are in east longitude from Paris, while Wallis is in west longitude, their dates differ by a day. Here are two volcanic islands; Fotuna, 40 km. in circuit and 850 m. in height, and Alofi, 20 km. in circuit and 200 m. in height. Both are singularly unlike Wallis in having strong slopes, rich forests that shade deep ravines drained by fine streams, only discontinuous fringing reefs instead of an encircling barrier reef, and therefore no good harbors. Viala describes these islands as "two pyramids, of which the flanks plunge into the sea in abrupt cliffs" (*falaises*); but the last term can hardly be correct, for the views of the islands on Hydrographic Office chart 1986 show the slopes to descend with almost even declivity from summit to shore. It may be added that neither the chart nor Viala's description suffices to determine whether the deep ravines lead down to embayments in the shore line or not; also, that in view of the presence of a number of submarine banks or "drowned atolls" in the north—a region without rival in this respect in the whole Pacific and for which the reviewer has therefore proposed the name "Darwin Hermatopelago" (*Bull. Geol. Soc. Amer.*, Vol. XXIX [1918], p. 531)—it is likely that the absence of a barrier reef here is to be explained by recent submergence at a rate too fast for reef upgrowth: hence whatever barrier had been formed around the two islands previous to this submergence should exist now as a submarine bench. The lack of soundings makes it impossible to test this supposition.

Viala gives interesting accounts of the natives, of whom there are 4,500 on Wallis and 1,500 on Fotuna and Alofi, with descriptions of their various customs, of the Catholic missions by which their mode of living has been much improved, and of the prevalent diseases. A noteworthy peculiarity of the natives is their boldness in risking inter-island voyages in their canoes, with only the rudest means of laying their course.

W. M. D.

The Reed-Wekusko Map-Area, Northern Manitoba. By F. J. ALCOCK. Ottawa: Canadian Geological Survey, Memoir 119, 1920. Pp. 47, pls. 6, maps 2.

The discovery of gold-bearing quartz veins and rich sulphide deposits in basic pre-Cambrian rocks of northern Manitoba has attracted con-

siderable attention. This memoir describes in some detail one of these areas of basic pre-Cambrian rocks and is representative of the geology of the mining camps of northern Manitoba.

The area lies along the border of the Laurentian Plateau to the north and the Great Plains to the west. The average elevation of the Laurentian Plateau part is about 950 feet above sea-level, the highest hill being 1,060 feet and the lowest flat 818 feet above sea-level. This hummocky surface of low relief represents the surface of a pre-Ordovician peneplain recently uncovered and slightly modified by Pleistocene glaciation. The streams are characterized by lake expansions, rapids, and waterfalls. Lakes with irregular outlines and many islands are abundant.

The rocks fall into four groups: (1) Pleistocene drift and stratified clay; (2) Ordovician dolomite; (3) pre-Cambrian granite and its differentiates; and (4) pregranitic complex of igneous and sedimentary rocks.

The Pleistocene deposits consist of drift, outwash material, and stratified clays deposited in glacial Lake Agassiz. The Ordovician dolomite forms an irregular escarpment across the southern border of the area. No clastic base is present and the Ordovician seas advanced over a slightly rolling surface. A few fossils of Trenton age have been found in this dolomite. The pre-Cambrian granites, the most abundant rocks of the area, are intruded as stocks and batholiths and vary considerably from place to place in both mineralogical and chemical composition. Massive reddish biotite or hornblende-biotite-granite is the most abundant type. In general these granites are massive, but in places gneissoid types occur. Pegmatite dikes and quartz veins represent the last phases of the intrusion. The pregranitic complex, the oldest rocks of the region, is divided into the Kiski volcanics and the Wekusko sedimentary series. The Kiski consists largely of volcanic rocks varying in composition from rhyolite to basalt. Beds of pyroclastics are found interbedded with the flows. These rocks are altered to sericite, hornblende, biotite, and chlorite schists. The sedimentary division or Wekusko series consists chiefly of garnet gneiss and mica schist with many other varieties of metamorphic sediments in smaller amounts. The series is of great thickness, is highly folded, and intensely metamorphosed. Except for a small area of slate, the series is coarse clastics which vary considerably in coarseness from place to place and with local conglomerate horizons.

The chief ore deposits of the region are gold-bearing quartz veins associated with the granite intrusives and cutting all pre-Cambrian

rocks of the region. These quartz veins are abundant along the contacts of the granite and the Wekuspo series. Some of the quartz contains gold in visible quantities. The veins are variable in width averaging from 18 inches to 12 feet and some of them have been traced for 1,600 feet or more. One carload of ore shipped from the Northern Manitoba group averaged $81.53 per ton in gold. Considerable development work on a number of the properties has already been done, and after transportation facilities are improved and mining conditions become normal, this should prove an important gold-producing region.

J. F. W.

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The Physical Chemistry of the Crystallization and Magmatic Differentiation of Igneous Rocks. IV. By J. H. L. Vogt.

Pleistocene Mollusca from Northwestern and Central Illinois. By Frank Collins Baker.

The Character of the Stratification of the Sediments in the Recent Delta of Fraser River, British Columbia, Canada. By W. A. Johnston.

Outline of Pleistocene History of the Mississippi Valley. By Frank Leverett.

Pennsylvanian Stratigraphy of North Central Texas. By Raymond C. Moore and Frederick B. Plummer.

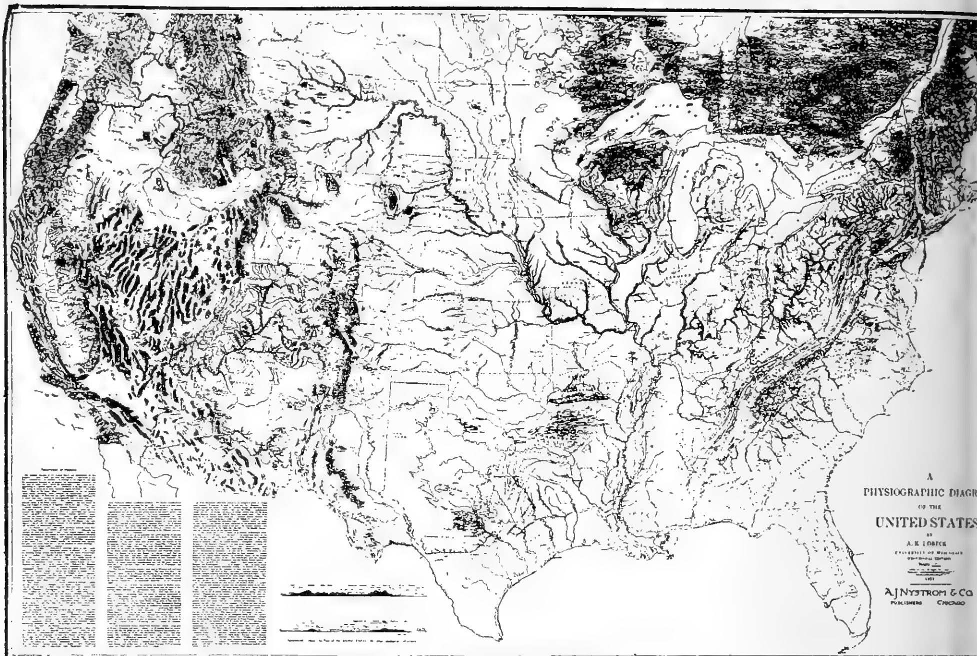
Suggestions as to the Description and Naming of Sedimentary Rocks. By A. J. Tieje.

The Marine Tertiary of the West Coast of the United States: Its Sequence, Paleogeography, and the Problems of Correlation. By Bruce L. Clark.

The Relation of the Physical Properties of Natural Glasses to Their Chemical Composition. By William O. George.

Notes on a Recent Collection of Fossil Plants from the Dalles Group of Oregon. By Ralph W. Chaney.

An Outline of the Geology of New Zealand. By W. N. Benson.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THEORETICAL CONSIDERATIONS OF THE GENESIS OF ORE DEPOSITS

R. H. RASTALL
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Nothing is more noticeable in the history of mining geology than the change which has taken place in the last half-century in the prevailing opinions as to the origin of ore deposits. We may leave out of account the absurdities of Wernerism and most of the earlier speculations of the Plutonist school. Scientific ideas as to the genesis of ores may be said to date from the publication of Breithaupt's epoch-making little book in 1849.[1] Here in the idea of paragenesis we have the germ of modern speculations on this subject. For long, however, the theory of the aqueous origin of all ore deposits held sway in its various forms, one of the best-known of which was the famous lateral-secretion theory. Even yet something very like this theory is held to furnish the best explanation of the origin of certain types, as for example the lead-zinc ores of the Mississippi Valley and other similar deposits. The brilliant writings of Pošepný also did much to perpetuate the reign of water. Nevertheless, at the present time no one will be found to deny that many and important ore deposits are of direct igneous origin. As a matter of fact igneous and aqueous origin are by no means incompatible. It may at once be admitted that watery solutions may be produced direct from magmas. Of this

[1] Breithaupt, *Die Paragenesis der Mineralien*, Freiberg, 1849.

we have abundant evidence in the steam that accompanies volcanic eruptions, and the geysers and hot springs belonging to the later stages of igneous activity in many parts of the world, such as the Yellowstone Park, Iceland, and New Zealand. At Steamboat Springs, Nevada, ores are now visibly being formed by hot springs which must derive their heat from intratelluric sources. It is natural to suppose, therefore, that many of the ancient ore deposits have been formed in a similar way. The common association of ores with propylitization and certain other types of rock-alteration points in the same direction.

Moreover, ore minerals occur in rocks undoubtedly igneous as well as in lodes and veins. Cassiterite is found as an original mineral in granites, in pegmatites, and in quartz veins in undoubted and visible connection with granites. The same applies to its constant companions wolframite and molybdenite. Gold is known in almost every conceivable geological situation of almost every age. Chromium and platinum are found beyond doubt as original constituents of ultrabasic rocks such as peridotites and serpentines. Magnetite and ilmenite have segregated in enormous masses from plutonic intrusions, and so on indefinitely. A great part of the rich mineralization of Mexico and South America is in obvious and visible connection with the great volcanic outbursts ranging from late Cretaceous to modern times. It is clear that a vast number of the most valuable ore deposits are of direct igneous origin. In many other instances, though the connection is not so clear, it is still highly probable or indeed certain.

Nevertheless, although the main fact in its broadest outlines is established, considerable doubt still remains as to the mechanism of the processes by which the concentration and deposition of the ores has been effected, and the underlying and fundamental reasons for these processes. Another highly important aspect of the subject is the distribution in space and time of the different types of mineralization and their relation to crust disturbances and various petrographic types. This in the broadest view is the scope of the study of metallogenesis.

Hitherto it has not been found possible to devise a really satisfactory and logical classification of ore deposits on a genetic

basis. This impossibility is inherent in the nature of the subject, since in nature there exist no sharply defined categories, no pigeonholes into one of which every type will fit. It is the intermediate and transitional types that are the bugbear of any such attempt. Nevertheless there are some basic facts that may be used as a groundwork for generalization. In the first place we have certain ores occurring as original minerals in igneous rocks, and as disseminations and magmatic segregations whose origin from magmas is beyond doubt. Also there are the innumerable instances of ore-bearing pegmatites which can also be assigned with safety to the same origin. A great number of contact and replacement deposits also undoubtedly owe their metal content to transfer from intrusive masses.

It is, however, when we come to the large and highly important class of deposits described in general terms as veins and lodes that difficulties begin to manifest themselves. These undoubtedly grade on the one hand into the magmatic deposits, while on the other hand some of them show distinct evidence of having been formed near the earth's surface at the ordinary temperature and pressure. In dealing with the doubtful members of this group we have to take into account not only the characters of the deposits themselves, but all the attendant circumstances which may throw any light on their origin, such as geographical distribution, relation to sedimentary and other formations of known age, and to the structure, disposition, and character of the surrounding rocks; in short, their geological features. It is the geology of the ore deposits that will throw most light on their origin.

As an example let us take the mining region of western Cornwall, one of the most highly mineralized districts of the world, especially as regards the number of metals found. Here cassiterite occurs as an original mineral in the granite, in pegmatites, in greisens and other pneumatolytic modifications of the granite, in quartz-porphyry dikes, and, most important of all, in a vast number of lodes and veins which chiefly congregate near the contact of granite and slate, almost invariably passing from one rock to the other without interruption. The lodes also show every possible degree of pneumatolytic alteration. Besides cassiterite they

contain wolframite, most abundantly in a comparatively narrow zone near the granite-slate contact. Arsenopyrite is also abundant, and in the upper parts of the lodes, farther away from the granite, tin gradually gives place to copper. Other metals, such as molybdenum, silver, antimony, bismuth, and uranium, are also found, while lead and zinc are abundant in a later series of lodes, usually more distant from the contact. Among the gangue minerals tourmaline, topaz, and fluorspar are abundant. Here the connection of the tin-copper mineralization with the granites and their pneumatolytic phase is obvious. The age of this is also definitely fixed, since the granites cut Upper Carboniferous rocks and the Permian strata are not metamorphosed. The whole of this igneous cycle is a direct consequence of the Armorican crust disturbances which had such an important influence in molding the geological structure of west-central Europe. Very similar phenomena are to be seen in Brittany, in Spain and Portugal, and in the Erzgebirge on both sides of the frontier between the German Republic and Czecho-Slovakia: all of these are broadly contemporaneous with the similar occurrences in Cornwall.

Here we have a clear example of a metallogenetic province, showing a definite association of mineral deposits of a peculiar type with a phase of igneous intrusion dependent on a particular set of earth movements. The number of metals present is very large, but the most characteristic are tin and tungsten, and a special feature observed in Cornwall, Bohemia, and Portugal is the presence of uranium.

Turning now to another region of the Old World, we find a great development of tin ores in the Malay peninsula, in Banka and Billiton, and on the eastern side of the Australian continent from Queensland to Tasmania. In Lower Burma (Tavoy) we find a little tin and much tungsten, so that this evidently forms a slightly varying extension of the same field, a local facies. This tin-bearing region stretches parallel to the great Malayan arc and its continuation into Australia and the mineralization is closely connected with the intrusion of granites, probably of Permo-Carboniferous age. Furthermore, this great province is divided up into subprovinces characterized by special mineral

associations, namely, in Tavoy, dominance of tungsten; in the Malay peninsula, Banka, and Billiton, dominance of tin alone; in Queensland, tin, tungsten, molybdenum, bismuth; in New South Wales, tin, tungsten, and molybdenum; in Tasmania, mainly tin.

The second most important tin-producing country of the world is Bolivia. Here along the Cordilleran chains, in association with the great Tertiary volcanic activity, we find extraordinarily rich veins carrying tin, tungsten, bismuth, and silver; a remarkable and apparently unique type.[1] These veins present features of great interest from the theoretical point of view. Three chief types of veins can be recognized, as follows:

a) Tin-bismuth veins with tourmaline and other pneumatolytic minerals, associated with deep-seated granites.

b) Tin-bismuth-silver veins, carrying most of the tin as complex sulphides, stannite, etc., associated with hypabyssal porphyry intrusions.

c) Silver veins without tin, associated with extrusive volcanic rocks.

A noteworthy feature is the occurrence of the tin in type (*b*) mainly as complex sulphides associated with argentiferous tetrahedrite and sulphosalts of copper, lead, bismuth, arsenic, and antimony. There is thus a distinct gradation in the metal contents of the veins, minerals with boron and fluorine occurring in quantity only in the high-temperature veins associated with granites. Thus the temperature-depth relations of tin and silver are clearly shown.

In all these cases there can be no possible doubt of the association of the tin and other metals with igneous rocks. When we turn to other metals the facts are equally striking. It is impossible to enumerate all of them, but a few examples must suffice, and these may be taken as typical of the rest. It has long been known that platinum has its original home for the most part in peridotites and their alteration product, serpentine, and to a less extent in certain nickel-bearing and other eruptives to be mentioned presently; in short, in ultrabasic and basic rocks. The peridotites also commonly contain large quantities of chromite and also chrome-bearing spinels: hence the presence of chromite is a useful

[1] Davy, *Econ. Geol.*, Vol. XV (1920), pp. 463–96.

indicator of the possible presence of the platinum metals. In the basic rocks also we find large masses of iron ore commonly rich in titanium, either as ilmenite or as titaniferous magnetite. The presence of titanium is specially characteristic of the basic group. Another type of great scientific interest and commercial importance is seen in the great segregations of nickeliferous sulphides found in connection with norites and gabbros, as at Sudbury, Ontario, and Insizwa in Griqualand East. Whatever may be the actual mechanism of the Sudbury nickel deposit, its close connection with the norite intrusion will hardly be denied, and at Insizwa the concentration of the sulphide by some form of differentiation seems clear. In association also with basic rocks, especially with gabbros, are found great masses of pyrite, as at Röros in Norway, and the cupriferous sulphides of the Huelva district (Rio Tinto and others) also belong here.

In all these cases of the occurrence of ores in basic magmas one point is worthy of special attention, namely, that the ores occur either as disseminations in the rock itself (platinum and chromium in part) or as segregations separated from the magma either by gravity, by diffusion to the margin, or by separation of immiscible liquids. The ore bodies therefore lie either within the intrusion as disseminations, schlieren, or irregular patches, or as a definite layer at its base, or as separate intrusions, often laccolithic in form. Here there is no separation of the ores in pegmatites and all the innumerable varieties of mineral veins, as with the acid rocks, while pneumatolytic effects are subordinate, or more commonly entirely absent. In conformity with this, metamorphic and metasomatic effects are also much less marked. The most characteristic type of basic pneumatolysis is scapolitization, which appears to correspond to tourmalinization in acid rocks, while the concentration of apatite in some instances may also be referred here.

Here another generalization of fundamental importance may be made: the ore segregations of the basic rocks are usually the *first* fraction of the magma to crystallize, those of the acid rocks are usually the *last* fraction. In neither case is the statement invariably true, but exceptions are as a rule due to special causes.

For example, the great iron-ore masses of Kiirunavaara and Luossavaara, as described by Daly, have been formed by the sinking of early-crystallized magnetite in a quartz-porphyry magma. The case of iron ore is, however, exceptional in that it is not specially characteristic of any particular magma, but is of more or less universal distribution.

Although this generalization must not be pushed too far, it is certainly true in its broad outlines. The reason for it is to be found in the presence in, or absence from, the crystallizing and differentiating magma of substances capable of combining with the metals to form volatile compounds, such as fluorine, boron, and especially water. We have every reason to believe that basic magmas do not contain much water; at any rate it is possible to produce basic rocks artificially from dry melts, whereas with acid rocks this cannot be done: some flux is required to insure the crystallization of quartz or orthoclase or hornblende. Such fluxes collect always in the acid fractions of a differentiate, there lowering freezing-points and in particular facilitating the concentration of metals in the last residue of the magmatic solution, which must always tend toward, if it does not actually reach, the composition of the multiple eutectic of the complex solution. In some simple cases the eutectic of quartz and felspar is actually reached, as seen from the simultaneous crystallization of quartz and felspar in graphic pegmatites.

The case of iron is a rather exceptional one, since it is found in considerable amount in connection with igneous rocks of both acid and basic types. The iron-ore masses of Sweden have already been mentioned: here the iron occurs as magnetite, occasionally as hematite, at all events as iron oxides only, and this is the common type in acid rocks. In the basic segregations, however, the state of affairs is generally different: in a great many ultrabasic rocks we find ferrous oxide combined with chromium as chromite, or with magnesium and aluminium in addition as picotite or some allied chromiferous spinel. In the felspathic basic rocks, on the other hand, the greater part of the iron is in some state of combination with titanium, either as ilmenite or as a titaniferous magnetite, whatever that may really be. Again, in the basic rocks iron is

found frequently as sulphide segregations, either with or without copper, nickel, and cobalt. Thus it is seen that although iron as an element is common to both ends of the series, nevertheless it occurs typically in a different state of combination in each.

Having thus arrived at the fundamental generalization that there are two principal groups of ore segregations of very early and very late consolidation respectively, we must now proceed to consider in more detail what particular type of differentiation may be applied in explanation of each.

At the present time five different sets of causes are commonly put forward in explanation of the varied set of phenomena comprised under the head of differentiation; including in this term not only the production of heterogeneity in single intrusions, but also the separation of minor magmas of varying composition from one original magmatic solution. These may be summarized as follows:

a) Marginal concentration of minerals of high freezing-point by diffusion of liquid molecules immediately preceding and during crystallization.

b) Differentiation by gravitational sinking or rising of crystals in a solidifying magma.

c) Separation of a magma into two or more immiscible or partially miscible layers.

d) Assimilation by stoping with its attendant fluxing effects.

e) Squeezing out of liquid residue from a crystalline sponge or network.

This is not the place to discuss the evidence as to the competence of each of these processes as a general factor in differentiation. It is probable that all are applicable in different cases and under different sets of conditions: we have only to consider which of them can be invoked to explain the various types of ore occurrence here already briefly alluded to. It is evident that in the two great groups of ores of early and late consolidation very different physical conditions must prevail, especially as regards temperature and presence of the so-called mineralizing agents or fluxes. In the early group the temperatures must be very high, and we have reason to believe that in the basic rocks fluxes are unimportant; hence it appears that any one of the first three categories should be specially applicable.

Of the first type we have an excellent example in the marginal phase of the gabbro of Carrock Fell, as described by Harker. Here a well-marked segregation of titaniferous iron ore is found along both steeply inclined margins of a laccolith of considerable size, believed to be still more or less in the same position as when intruded. From the figures and description given by Harker it is clear that gravity is excluded, and the only possible explanation is some kind of diffusion to the cooling surfaces during crystallization. The concentration is not very strongly marked, as the most concentrated type of ilmenite-gabbro has only about 25 per cent of iron ore, but at any rate the principle is clear. A very peculiar type is the great mass of titaniferous iron ore at Taberg in Sweden, which forms the central portion of a boss of ultrabasic rock, usually described as hyperite, that is, olivine-norite in modern terms. The reason for this reversal of the normal sequence is unknown.

It will doubtless be generally conceded that it must be difficult on field evidence alone to distinguish between cases of heterogeneity due to gravity-sinking and to immiscibility in the liquid state. In both cases, on complete solidification the heavier rock will be found below, and owing to the high viscosity of fused silicates no very sharp line of demarcation is likely to be seen. It is moreover highly probable that in some instances both causes have been operative. Bowen has brought forward much evidence in support of differentiation by gravity-sinking, and Vogt long ago showed that fused silicates and sulphides possess very strictly limited miscibility.[1] In the well-known instances of Sudbury and Insizwa we have thick intrusions ranging from acid or intermediate at the top to basic or ultrabasic below, with a well-marked layer of sulphides at the bottom. Here it may be suggested that the gradation in the silicate rock is due to gravity with an immiscible separation of sulphide at the base. It seems pretty clear from Daly's latest observations that the concentration of magnetite and apatite at Kiruna is due to gravity-settling of rather large units of differentiation, since many blocks of ore have been caught and fixed at higher levels by increasing viscosity. The occurrence of limited miscibility in silicate solutions is a much disputed point,

[1] Vogt, *Die Silikatschmelzlösungen*, Part I (Kristiania, 1903), p. 96; "Die Sulfid-Silikatschmelzlösungen," *Norsk geologisk tidsskrift*, 1917.

as to which we have as yet no absolute evidence, but it seems quite clear that this cause may be legitimately invoked to explain cases of magmatic sulphide segregations. Vogt has shown by quantitative determinations that the mutual solubility of norite magmas and sulphides is very small indeed; at 1300° C. and one atmosphere, less than 0.5 per cent in the case of pyrrhotite, while copper and nickel sulphides are practically insoluble in norite melts. Moreover, it is improbable that these relations are seriously altered by high pressures.[1]

Up till the present time we have no definite information as to the part played by assimilation and stoping in the formation of ore deposits. In the nature of things such a process would necessarily be difficult to detect, since it would itself destroy its own traces. It is possible, however, that some masses of ore in acid and intermediate rocks may have been introduced in this way, such as the deposits of iron, copper, and gold in magmas of the monzonitic and dioritic facies. This, however, is pure speculation with no tangible evidence to support it. It seems possible that in the earliest solid crust of the earth, afterward remelted, there may have been segregations of metallic ores, afterward reabsorbed and differentiated, as suggested by Morrow Campbell.[2] As to this, likewise, there can in the nature of things be no positive information. It is known that in some localities, for example, southwestern Norway, there is a notable concentration of minerals of the rare earths into syenitic pegmatites, as described by Brögger.[3] If we accept Daly's theory of the origin of alkaline rock types by the fluxing effects of assimilated limestone in normal magmas, it would seem to follow that the concentration of the rare earths here must also have resulted directly from the assimilation. On this question as a whole further information is needed, as few observations are available on the occurrence of ores in connection with highly alkaline magmas.

With regard to the squeezing out of liquid residues from partially consolidated rocks, like water from a sponge, as so graphically

[1] Vogt, *Norsk geol.tidsskr.*, 1917, p. 77.

[2] *Bull. Inst. Min. Met.*, October, 1920, p. 3.

[3] Brögger, *Zeitschr. für Kryst.*, Vol. XVI (1890).

pictured by Harker,[1] this cause may undoubtedly be invoked to explain the separation of pegmatitic material from cooling granites under earth pressure. This process has given rise to areas of pegmatitic permeation, as described by Barrow,[2] in the south-eastern Highlands of Scotland. It is hardly necessary to point out again that pegmatitic mother-liquors often carry notable amounts of ore minerals, and in many instances this seems the most reasonable explanation of ore-bearing pegmatites.

From the foregoing brief and imperfect summary it appears that ore deposits may arise from all or any of the physicochemical processes that have been invoked to explain the origin of heterogeneity in igneous rocks. In some cases one is applicable, in some cases another, but none are apparently excluded on a priori grounds. This variety of origins is only what we should expect from the great differences observable in the characters of the deposits themselves. Here again a correlation can be traced between the nature of the effective process and the physical properties of the molten magmas of different composition and especially the degree of viscosity. Generally speaking, it is admitted that basic magmas are more liquid than normal acid magmas in spite of the higher proportion of water in the latter. Hence differentiation by diffusion, gravity, and immiscible separation are more marked in the basic class, owing to lower viscosity. It is only in the differentiated and concentrated last residues of the acid class that the solution becomes readily mobile and lends itself to formation of veins and pegmatitic permeation. A similar effect is produced by assimilated fluxes in certain cases. These considerations lead again to the same conclusion as before, namely, that the ore deposits of the basic rocks are marginal and basal, or included segregations of a streaky nature if convection currents have been active, while from the acid fluxed magmas are formed veins and lodes in all their varieties, either in fissures in the rock itself, often congregated near its roof, especially in subsidiary domes and cupolas,[3] or actually external to the intrusion, filling fault planes and other fissures and zones of weakness in the

[1] Harker, *Natural History of Igneous Rocks* (1909), p. 323.

[2] Barrow, *Quart. Jour. Geol. Soc.*, Vol. XLIX (1893), p. 330.

[3] Butler, *Econ. Geol.*, Vol. X (1915), pp. 101–22.

country rock. The endless varieties of contact deposits are not here taken into account in detail, but it may perhaps be said that they are more characteristic of acid intrusions, doubtless owing to the abundance of mineralizers, which act as carriers for the metals.

Having thus considered in general terms the phenomena of differentiation and concentration in igneous rocks in their bearing on the origin of ore deposits, it remains to see whether it is possible to draw up a schematic classification on this basis. If this can be done it would constitute a genetic classification of ore deposits in the strictest sense of the word; such a classification, if presented in a graphic form, may be regarded as a genealogical tree of the ore deposits, or rather of the metals, as for this purpose it is hardly necessary to take into account their state of combination; furthermore it is obvious that the characters and distribution of the great class of secondary deposits have no bearing on the point: it is with primary ores alone that we are concerned. Now most primary ores are either native metals, oxides, or sulphides—in some instances arsenic also seems to play the part of an electronegative element, as in enargite and several nickel and cobalt minerals and in arsenopyrite. Furthermore, most primary ores are of very simple composition: the more complex minerals are mostly found in the oxidized ores and in those of the zone of secondary enrichment, with which we are not here concerned.

The primary facts that we have to start on are somewhat as follows:

1. It has been shown that certain metals and non-metallic elements are found chiefly in the basic rocks, including especially nickel, cobalt, chromium, platinum, titanium, with chlorine, phosphorus, vanadium, and sulphur. Iron sulphides, often nickeliferous, are common.

2. Of these, chromium and platinum are specially characteristic of the ultrabasic rocks, themselves usually differentiates of basic magmas.

3. In connection with the extreme differentiation phases of acid magmas are found especially tin, tungsten, molybdenum, zirconium, uranium, lithium, with fluorine, boron, and abundance of water.

4. The rare earths of the thorium-cerium group are found chiefly in pegmatitic phases of granitic and syenitic magmas.

5. Iron, gold, silver, and copper are of more general distribution, without much tendency to characterize any special type of magma. Iron, however, shows a difference of combination: in the basic rocks it is largely combined with sulphur, titanium, or chromium, in the acid rocks it concentrates mainly as magnetite. Primarily copper is generally combined with sulphur, less commonly with arsenic, wherever it is found.

6. It appears that in the earliest stages of earth history known to us there were two primary magmas, granitic and basaltic: from these the other existing rock types were derived by differentiation.

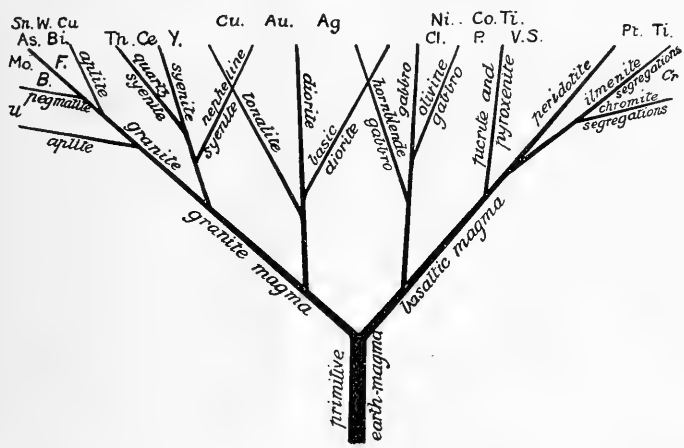

FIG. 1.—Genealogical tree illustrating conceptions of differentiation of igneous rocks and ore deposits from primitive magma.

It is a fair inference that earlier still there was one primitive earth magma of intermediate composition, perhaps a siliceous immiscible fraction, separated from a mainly metallic fraction that formed the heavy core of the earth. In this primitive silicate magma all the valuable metals were at first in solution, afterward separating into its various differentiated fractions according to their partition coefficients, or their degrees of solubility—no doubt assisted by gravity-sinking and fractional crystallization. Some metals seem to have been more or less soluble in all magmas, while others had very decided preferences for a particular type.

It is on the basis of these principles that the genealogical tree here presented (Fig. 1) has been constructed. In its essential

features it is similar to a diagram already published by the writer.[1] The present form of it is, however, slightly more elaborate and the various branches have been arranged in accordance with the silica percentage of the rocks, although the figure must not be regarded as drawn strictly to scale. It is also to be remarked that the thickness and length of the branches do not bear any relation to the actual abundance of the different rock types: in this respect the figure is purely diagrammatic and not quantitative. The angles of inclination between the various branches are controlled only by convenience of drawing. The intercrossing of various branches is, however, intentional, in order to show that similar final products may be obtained from different partial magmas. That is to say that similar rocks and ore types may have different chains of descent from a common ancestor. This is analogous to the biological phenomenon known as heterogeneous homoeomorphy.

This diagram, when looked at from this point of view, is in fact a basis for a truly genetic classification of certain large groups of primary ores; nevertheless it must be regarded as strictly limited in its scope. It takes no account of the secondary ores, or of certain groups which may be of supergene origin, formed by the action of descending meteoric waters. With these classes we are not now concerned. But it is claimed that this systematic arrangement does throw some light on the origin and genetic relationship of the great classes of primary ores, which are of fundamental importance, as being in all probability the original source of all the workable metallic deposits of the globe.

In this treatment of the subject it is necessary to take into account also the conceptions of metallogenetic epochs and metallogenetic provinces, which have been so ably worked out of late years by many writers, especially in America. If the whole scheme of differentiation as here outlined is regarded as continuous, a false impression will be obtained. On the contrary, the processes are discontinuous both in time and space. An admirable summary of the present state of our knowledge of the chronology of ore

[1] Rastall, "Differentiation and Ore-Deposits," *Geological Magazine*, Vol. LVII (1920), p. 298.

deposits will be found in the last chapter of the new edition of Lindgren's *Mineral Deposits*. From this it is clear that the type of mineralization here considered is mainly restricted in time to three periods, pre-Cambrian, Permo-Carboniferous, and Tertiary, corresponding to the greater periods of crustal disturbance and mountain-building. Of these the first is certainly complex and of great length, probably including several periods comparable in magnitude to the two later ones, but as yet it is hardly possible to disentangle these clearly in most parts of the world. The Canadian shield is an exception.[1] Here several distinct periods are distinguishable. Nevertheless it must be remembered that although the now visible manifestations, including the actual deposition of the ores, were spasmodic, the genetic processes are undoubtedly in constant operation in depth, elaborating the material and preparing the way for the igneous outbreaks and accompanying ore-formation that occur whenever the static equilibrium of the crust is disturbed, whatever may be the cause to which these disturbances are due. This is a fundamental question into which we cannot now enter.

[1] See Miller and Knight, *Trans. Roy. Soc. Canada*, Vol. IX (1915), pp. 241–49.

NOTE ON A POSSIBLE FACTOR IN CHANGES OF GEOLOGICAL CLIMATE

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In a recent discussion of the factors that control variations of world climatic conditions, Professor Humphreys has criticized the various astronomical hypotheses that attempt to explain the ice ages and other features in geological climate.[1] Several factors, formerly given weight, are held responsible only for effects of the second order at most. Among the insufficient interpretations he would place Croll's theories which involve the changing elements of the earth's orbit, and the various hypotheses connecting sun spots (and other intrinsic changes in solar radiation) with terrestrial insolation and the climates of the geologic past.

The primary factors of climatic control clearly appear to be of terrestrial origin—land elevation (and its concomitant factors of oceanic and atmospheric distribution and circulation), combined with vulcanism. Cosmic factors are evidently not of primary importance. The observed connection of climatic phenomena with land elevation can by no means be attributed to chance. A cosmic non-terrestrial origin for the principal factors of climatic control would therefore leave unexplained such significant coincidences as mountain forming and glaciation, since one could hardly propose seriously that cosmic factors are the cause of both topographic and climatic change.

One astronomical possibility that I believe has not been urged heretofore appears, however, as a result of recent observation, to deserve attention as a potential secondary factor in recorded geological climates, and possibly even as a primary agent at some prehistoric time.

[1] W. J. Humphreys, "Factors of Climatic Control," *Jour. Frank. Inst.* *CLXXXVIII* (1919), 775; *CLXXXVIII* (1920), 63.

1. Barnard, Wolf, and others have shown the very common existence of extensive, diffuse, irregular, luminous and non-luminous nebulosities in the sky, and I have previously called attention to the apparent affiliation of Barnard's dark nebulae with our local cloud of stars; Hubble's unpublished observations appear to substantiate this affiliation of the dark markings and demonstrate their relative proximity to the solar system.

2. The great Orion nebula is one of the luminous or partially luminous members of this group of nebulosities, and spectroscopic observations by Fabry, Frost, and Campbell and their associates have shown that its various parts are moving relative to each other with velocities of several kilometers a second.

3. In and near the Orion nebula more than seventy faint stars that vary in light have been found at the Harvard, Heidelberg, and Yerkes observatories, and elsewhere; Lampland has also found similar variables in the comparable nebula Messier 8. My work on the variables in Orion has shown: (*a*) no certain periodicity in the variation, (*b*) light curves not comparable with known types, (*c*) a variety of color types among the variables, and (*d*) no evidence of great range or of extinction (as would be expected from occultations).

4. Van Maanen's discussion of the proper motions supports the inference, based on the distribution of the variables, that they are really associated with the Orion nebula and with the cluster of brighter stars in that vicinity. The distance of this group of stars, according to Russell and Kapteyn, is six hundred light-years. The diameter of the region throughout which these peculiar variables are known is fifty thousand times the diameter of Neptune's orbit, and the total nebulous region in Orion is many times larger. (From the present direction and amount of their motions, we compute that a few million years ago our sun was in the vicinity of the Orion nebula; at its present speed the sun would require nearly a million years to pass through that particular nebulous region).

5. From the known distance of the variables in Orion, and my measures of their apparent magnitudes, it is easily computed that in luminosity they are dwarfs. This supports the conclusion from the study of their light curves that the Orion variables differ

from all other types of variables, which almost without exception are giants in luminosity.

6. In view of (*a*) the irregularities of the light variations, (*b*) the apparent immersion of the variables in nebulosity, and (*c*) the spectroscopic evidence for the irregular churning about of this nebulous matter, it seems reasonable to believe that the variations in brightness result from collision or friction with the irregular nebulosity in which the variables are involved. The encounter of star with nebula is at present the best, though perhaps not an entirely satisfactory, explanation of the cause of galactic novae; and it is the only hypothesis that has been suggested to account for temporary stars in the rapidly moving spiral nebulae.

7. Long-exposure spectrograms of the Orion nebula, using the 100-inch reflector and a rapid focal-plane spectrograph, have recently shown in the bright-line spectrum the presence of hydrogen, nebulium, helium, carbon, and nitrogen; they also show a faint continuous spectrum in all parts of the nebula.

8. The bearing of the foregoing observations on the question of geological climates becomes obvious when we note the following points: (*a*) The condition that causes variation of a star in the Orion nebula must also gravely affect the atmosphere surrounding any attendant planet. (*b*) The sun is moving with a velocity of 20 kilometers a second through a region of space, large sections of which are known to be occupied by diffuse nebulosity (most of it probably much less dense than that in Orion). (*c*) The observed variation of the friction variables in Orion is generally from 20 to 80 per cent of the total light, and sometimes appears rapidly oscillatory, sometimes secularly progressive, sometimes a discontinuous brightening or dimming. (*d*) A change of 20 per cent in the solar radiation, if maintained for a considerable period of years, would sufficiently alter terrestrial temperature to bring on or remove an ice sheet; an 80 per cent change, unless counteracted by concurrent changes in the terrestrial atmosphere, would completely desiccate or congeal the surface of the earth.

October, 1920

THE PLEISTOCENE SUCCESSION NEAR ALTON, ILLINOIS, AND THE AGE OF THE MAMMALIAN FOSSIL FAUNA

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Some time previous to 1883 the late Honorable William McAdams, an ardent naturalist of Alton, Illinois, collected a number of mammalian remains from the Quaternary deposits of the Mississippi bluffs near his home city. The specimens found their way to the United States National Museum at Washington, where they have been studied by Dr. O. P. Hay. The collection includes remains of a ground sloth, horse, peccary, a large deer, moose, reindeer, eland, musk ox, mastodon, beaver, ground hog, pouched gopher, and brown bear.[1] Their importance became obvious when some of the individuals were found to represent new species, and therefore it was especially desirable to ascertain if possible their stratigraphic horizon.

The attention of the writer having been called to this collection through the kindness of Doctor Hay, arrangements were made with the Illinois Geological Survey for a few days' study with the hope of finding the exact locality and horizon, and of determining the place of the deposits in the Quaternary series. The deposits proved to be so suggestive of new interpretations that this preliminary paper has been prepared.

Previous literature.—In 1882 A. H. Worthen noted the finding of a portion of the jawbone of a mastodon in the lower part of the loess just northwest of the city of Alton.[2] In 1883 McAdams published an abstract of a paper in the *Proceedings of the American*

[1] O. P. Hay, personal communication.

[2] A. H. Worthen, "Geology of Madison County," *Economical Geology of Illinois*, Vol. I (1882), p. 252.

Association for the Advancement of Science which he read at the Minneapolis meeting. Quoting from this paper[1] we read:

> The drift clays proper at Alton, Illinois, had a maximum thickness of about one hundred feet, and the bluff clays were nearly of the same thickness. These clays were remarkably rich in animal remains, such as teeth and bones, attached to calcareous nodules or claystones. Remains of thirteen different species, now perhaps all extinct, had been found. The rodents were well represented by bones of seven species, including three or more beavers and some gophers. Nearly seventy teeth were found in the Quaternary deposits, a majority of them in a single quarry.

The locality.—The bluffs northwest of Alton have a height of from 125 to 175 feet above the Mississippi River. Approximately 75 to 125 feet of the section is Mississippian limestone, while the overlying material consists largely of loess with a thin deposit of glacial till in places separating the loess from the limestone. The upper surface of the limestone is somewhat uneven, but the cliffs persist for many miles up the river. Just east of Alton the Mississippian formations give way to the Pennsylvanian sediments. The upland of the western part of the city is characterized by karst topography, due to the solvent action of ground water on the underlying Mississippian limestone.

Section at one of the quarries.—At Plant No. 2 of the Mississippi Lime and Material Company, northwest of the roundhouse, the following section was found:

	Feet
Soil, loessial, dark brown, leached	1
Loess, brownish at the top, grading below within a few feet into buff; leached 4 to 5 feet below the soil; calcareous below, with a few loess fossils scattered through the deposit; stands with steep face in fresh cuts; maximum thickness . . . . about	20
Loess distinctly more reddish than the overlying, strongly suggesting a distinct deposit; contains many fossil snails, some "pipestem" concretions; has a silty texture but not sandy; maximum thickness . . about	30
Glacial till, reddish, contains many erratic pebbles of granite, dolerite, greenstone, quartzite, and other Canadian rocks, and also some local rocks, mostly subangular, some	

[1] *Proc. A.A.A.S.*, Vol. XXXII (1883), pp. 268–69.

rounded. The till is much more oxidized than the overlying loess. In the contact zone between the till and the loess are the lime concretions from which the mammalian remains were obtained.[1] Thickness . . .	1–3
Limestone, Mississippian age, about	100

Salient points regarding the drift.—As noted above, the glacial till exposed at this quarry is very thin and much more oxidized than the overlying loess. At the next quarry to the northwest, a patch of glacial till about 12 feet thick lies beneath the loess, strikingly set off from the loess above by a chocolate brown, weathered zone. The face of the exposure above the limestone cliff was too precipitous to admit of detailed study, but in the succeeding quarry to the northwest, at Plant No. 3 of the Mississippi Lime and Material Company, near the northeast end of the quarry, 18 feet of till above the limestone and below fossiliferous loess was accessible for study.

The till is a typical pebbly clay till. Its top is rounded and marked by a pebble band. The till is also oxidized to dark brown, changing below within a few feet to a yellowish brown to yellow color. The matrix of the till and the limestone pebbles are leached for about 8 feet down from the summit of the till, but where the slope of the till cuts down to a lower level this leached zone is diminished to about 3 feet. Beneath, the till is highly calcareous, and contains lime concretions up to about 8 inches in diameter. No such concretions occur in the overlying loess. The drift breaks into small polyhedral forms, due to the numerous intersecting joints. The surfaces of these joints are coated with oxide of manganese.

Age of the drift.—That the drift is much older than the loess is clearly shown by the weathered zone at its top. Its geographic location brings it within the possibility of being Illinoian in age; but there are strong suggestions that it may be as old as the Kansan. The 8-foot depth of the leached zone beneath fossiliferous and calcareous loess, the chocolate brown color of the upper part of the oxidized zone, the coating of manganese dioxide along the numerous joints which transect the till body, and the large concretions

[1] Fortunately the writer was able to have Mr. John D. McAdams, son of the collector, verify the horizon and the concretions as the source of the fossils.

in the calcareous zone, representing a concentration of part of the lime formerly in the upper portion, are somewhat extreme for drift of Illinoian age, underlying calcareous loess. The significance of these evidences seems even greater when one takes note of the pebble band at the top and the rounded summit which indicate the removal of an unknown amount of the leached drift. That this erosion was considerable and took place before the deposition of the loess is further suggested by the patchy distribution of the drift.

The evidences of great age stimulate a search for the nearest known Kansan drift. Drushell describes several exposures of much weathered drift, believed to be Kansan, in St. Louis.[1] And Fenneman states that the vicinity of St. Louis "apparently contains the thin edge of the Kansan drift sheet," as well as that of the Illinoian drift.[2] The Kansan drift of northern Missouri being in the Keewatin field, the question arises as to whether the remnants of drift at Alton may not record the movement of the Kansan ice across the site of the present Mississippi Valley. An examination of the petrology of the Alton deposits, made with the capable help of Dr. T. T. Quirke, supports an affirmative answer. None of the pebbles, so far as known, is distinctively from ledges in the Labrador field, whereas, all of them may well have come from formations in the Keewatin field, a few strongly suggesting native ledges at the west end of Lake Superior.

Some negative evidence, however, is to be considered. Leverett has observed two occurrences of glacial striae on bedrock at Alton, whose trend is S. 30°–40° W.[3] Although referred to the Illinoian ice, they were probably made by whatever glacier deposited the overlying till. If the drift is Kansan and of Keewatin relationship, the ice lobe in northern Missouri must have radiated until the eastern peripheral part moved in a northeasterly direction (E. 50°–60° N.)—a seemingly extreme departure from the main movement

[1] *Jour. Geol.*, Vol. XVI (1908), pp. 493–98.

[2] "Geology and Mineral Resources of the St. Louis Quadrangle," *U.S. Geol. Survey*, Bulletin 438, p. 31.

[3] Frank Leverett, "The Illinois Glacial Lobe," *U.S. Geol. Survey*, Monograph XXXVIII (1899), pp. 86–87.

of the lobe. Hence, from this standpoint, it seems better to credit the striæ to an ice sheet coming from the northeast, and to inquire into this possibility.

Pre-Illinoian drift in the Labrador field has been identified in surface exposures in the western part of the La Salle quadrangle and the eastern part of the Hennepin quadrangle.[1] Drift below the Illinoian has also been reported in a recent well in the Springfield region.[2] Such drift is far enough east of the known limits of the Keewatin field to be best referred to the work of a glacier in the Labrador field having a general southwesterly movement in Illinois. But, as pointed out above, the materials collected from the Alton deposits do not show anything distinctively Labradorean in origin. More data must be awaited to settle definitely the age of the drift concerned.

The loess deposits.—As given in the section described above there are two deposits of loess in this region, separable at least on the basis of color. The lower loess is characterized by its reddish tinge, the upper by its buff color. At all of the fresh exposures of the several quarries at Alton, this difference is conspicuous, and this is found to hold not only for street cuts in other parts of the city but for more distant localities. Buff loess overlying reddish loess is shown at the quarries at Grafton, 18 miles up the river, and at Edgewood, about 25 miles down the river. At Alton the reddish loess is somewhat thicker than the buff, the former being about 30 feet thick, the latter about 22 feet. Approximately the same ratio holds true at Grafton, but at Edgewood the conditions of slope do not favor a good estimate. Both are fossiliferous, but on the whole the lower seems to contain more shells than the upper. The fossil content will be discussed later. In texture, both seem to be typical loess, with no notable quantity of sand or suggestion of stratification.

Both are doubtless aeolian in origin, their relations to the Mississippi Valley indicating its flood plain as the chief source of

[1] Gilbert H. Cady, "Geology and Mineral Resources of the Hennepin and La Salle Quadrangles," *Bulletin 37* (1919), p. 71.

[2] E. W. Shaw, and T. E. Savage, "The Tallula-Springfield Quadrangles," *U.S. Geol. Survey*, Folio No. 188 (1913), p. 7.

supply. The distinctness between the two in color remains to be explained. Three working hypotheses may be tested. (1) The difference in color is original, the lower having been derived from a reddish silt, the upper from a buff silt. (2) The difference in color was brought about by weathering of the lower loess before the upper was deposited. (3) The lower was deposited under climatic conditions which favored oxidation to a reddish tinge during accumulation and then with a change in climate the succeeding deposits were not so highly oxidized, only a buff color being produced. The first and third hypotheses do not necessarily imply a distinct interval between the two stages of deposition, while the second, of course, does. A solution was sought in the nature of the contact, the evidences of weathering, and the fossil content.

1. *Character of the contact.*—In perhaps most places the contact between the two loesses is gradational, the gradation, however, taking place within a few inches. At first glance, this might be taken to record continuous deposition. But such an interpretation must not be held too rigidly in the case of wind deposits, since materials of an old surface may be mixed with new sediments, either by the wind itself or by organic agencies, thereby obscuring the record of an interval.

2. *Difference in oxidation.*—The reddish tinge of the lower loess is somewhat greater at the top than lower down in this deposit, although the reddish tinge characterizes the whole deposit. The stronger color at the top of the lower loess suggests an interval of weathering before the deposition of the overlying, but the reddish tinge throughout must be otherwise explained.

3. *Differences in content of lime carbonate.*—If the lower loess was subjected to weathering, it should show evidence of leaching of lime carbonate. The acid test revealed the presence of some lime carbonate throughout the reddish loess and into the overlying buff loess, but the reaction was notably feebler in the top of the reddish loess than in the base of the buff. While not strong, this evidence is suggestive of at least a brief interval of weathering between the two loess epochs although it is realized that there may have been an original difference in the lime content of the two loesses. Some of the larger shells, such as the *Polygyras* which extend from bottom

to top in the lower loess, are more fragile in the upper part, thus strengthening the evidence of leaching.

4. *Bearing of the fossil content.*—A small collection of fossils from each of the two loesses was made, and these have been examined by Curator Frank C. Baker of the Museum of Natural History, University of Illinois. The fact that the collection was not exhaustive limits the conclusions that may be drawn as to the fossil differences between the two deposits. However, two points seem to be clear: first, the lower loess has notably more fossils than the upper; and secondly, the large *Polygyras*, so characteristic of the lower deposit, do not seem to pass into the upper. Two of these *Polygyras* have been designated new varieties by Professor Baker; one, *Polygyra profunda pleistocenica*, is smaller than the normal species of today, and the other, *Polygyra multilineata altonensis*, is larger than its living representative. These and some of the other forms are described in a paper by Professor Baker.[1] If these should be found to be limited to the lower loess, the fact might indicate some climatic condition of unusual character.

From the foregoing evidence it would seem that a brief epoch intervened between the deposition of the reddish loess and the buff loess, but that the difference in color is not wholly due to this epoch of weathering. There is the possibility that the reddish tinge was given the lower loess by oxidation during the period of accumulation, but this would imply a climate of rather moist and dry extremes. The helices of the lower loess, such as *Polygyra profunda* and *multilineata*, according to Baker, are fond of relatively damp places in forest litter, beneath old logs and rubbish, and hence seem to be negative evidence. However, the *Polygyras* of this deposit have variational differences from these types, and these differences may be significant in this respect, although this is conjecture. The alternative view that the reddish tinge is due to the original color of the silts which were then present on the Mississippi flood plain, is somewhat favored. Reddish silts beneath loess are known upstream at least on the Iowa side where they are described as "red loam."[2] The writer has also observed "slack-

[1] *Nautilus*, Vol. XXXIV (1920), pp. 61–66.

[2] W. H. Norton, "Geology of Scott County," *Iowa Geol. Survey*, Vol. IX (1899), pp. 486–88.

water" silts of various colors, including maroon, in a tributary to the Mississippi River near Clinton, Iowa. Although these silts are believed to be younger than the "red loam," they help to show that the Mississippi has carried reddish silts at different times.

The age of the loesses.—The occurrence of the loess on the weathered till shows that for a long time after the deposition of the glacial till there was no loess in this vicinity. If the drift is Kansan in age, the reddish loess may be Sangamon; if, on the other hand, the drift be Illinoian, the reddish loess probably is Peorian. It is unlike any Peorian loess of which the writer knows, but the color does not necessarily preclude that possibility.

The upper buff loess is leached to a depth of 5–6 feet from the surface—an amount not exceeding some instances of leaching of loess on the Early Wisconsin. However, the summit is somewhat rounded, favoring some slope-wash, and hence this figure probably does not represent the total amount of leaching. According to Curator Baker, one species (*Pyramidula shimekii*), typical of the Early Peorian loess,[1] occurs in the collection taken from the buff loess but not in that from the reddish loess. The thickness is unusual for post-Wisconsin loess, but its proximity to the Mississippi makes this possibility plausible. Thus, the evidence in hand is not decisive as to whether the upper buff loess is Peorian or Early Wisconsin. If the reddish loess is Sangamon in age, the absence of a record of a long interval between its deposition and that of the overlying buff loess would seem to tie the two rather closely and favor the Early Peorian age of the upper loess.

The mammalian fossil horizon.—Mr. McAdams reported the mammalian fossils to be associated with calcareous concretions in the lower part of the loess.[2] They occur, however, at the base of the loess and the top of the till. Scores of concretions were broken open and examined in the hopes of finding more fossil remains, but reward was had in but one specimen, which contained a remnant of the lower portion of some tooth of an unknown mammal. A quarryman at Plant No. 2 reported that in the summer of 1919 a

[1] Formerly called Iowan loess, but recently pointed out as Early Peorian in age: "The Iowan Drift a Review of the Evidences of the Iowan Stage of Glaciation,," *Iowa Geol. Survey*, Vol. XXVI (1917), pp. 140–64.

[2] *Proc. A.A.A.S.*, Vol. XXXII (1883), p. 268.

fragment of a jawbone with teeth was found attached to a concretion when the overlying clay was being washed away by the hydraulic method. Unfortunately, the specimen was not saved.

The concretions range in size commonly from 1 inch or less to 5 and 6 inches in their longest diameter. Some include *Polygyra* shells, characteristic of the lower loess from bottom to top, while others contain pebbles of the underlying drift. A large percentage of each concretion appears to be of very fine material, which has given them the name of "loess nodules," although they do not occur strictly in the loess. They are secondary, younger than both drift and loess, their content of lime having been dissolved from the overlying formations which, together with the finely divided material of the loess carried probably in the colloidal state, has been deposited in concretionary form in the contact zone between the loess and the drift. In the forming of some of these concretions some of the pebbles in this zone were included, and, in some instances, mammalian remains.

Composition of the fauna and its age.—The following is a list of species which have been identified by Dr. Hay[1] in the McAdams collection:

Megalonyx jeffersonii	Extinct ground sloth
Equus sp. indet.	Extinct horse
Platygonus cumberlandensis?	Extinct peccary
Sangamona fugitiva	Large extinct deer
Cervales roosevelti?	Extinct moose
Rangifer muscatinensis?	Extinct reindeer
Taurotragus americanus	Extinct American eland
Symbos promptus	Extinct musk ox
Mamut americanum	Extinct mastodon
Castor canadensis	Canadian beaver
Castoroides ohionensis	Extinct giant beaver
Marmota monax	Ground hog
Geomys bursarius	Pouched gopher
Ursus americanus	Brown bear

The quarry from which most of the remains were secured shows a gentle depression at the top of the limestone, as if the surface of the limestone led to a sink, such as characterizes the karst topography of the upland directly back from the quarries. An examination of two of these sinks revealed their origin to be joint

[1] Personal communication.

planes widened by solution. At the quarry, several solution channels are well exhibited and the faces of the rock along the joints are in some instances coated with travertine, and in others with silt or clay which has been carried down from the surface. The clogging of these subterranean channels would give rise to ponds in the surface sinks, modern examples of which occur a short distance north of Plant No. 3 of the Mississippi Lime and Material Company. Such situations may have existed during the Pleistocene, and if so, offered favorable conditions for the entrapment of the region's fauna of both water and land species.

The fossil fauna, according to Baker, contains certain elements which suggest the arctic phase of a glacial epoch and other elements which belong to the warm phase of an interglacial epoch. The remains, however, could not have lain on the drift during the long period of weathering which followed the deposition of the drift. Hence, it is thought the remains of the arctic elements are a record of the life which lived in the latter part of the next glacial epoch, while the remains of the warm fauna represent the life which lived at the beginning of the interglacial epoch following that glacial epoch. As the melting of the ice sheet was a response to the change of climate from cold to warm, it is believed that with the local disappearance of the ice sheet, the warm interglacial fauna succeeded the arctic before the deposition of the reddish loess.

If the till proves to be Kansan in age, the weathering of the drift may be credited to the Yarmouth interglacial epoch, the mammalian fauna to late Illinoian and early Sangamon times, the reddish loess probably to the Sangamon, and the buff loess to the Iowan. If, however, the till is Illinoian, then the fauna probably is a partial record of the life of the latter part of the Iowan glacial epoch and the early Peorian interglacial epoch. In the latter case, the writer finds some difficulty in assigning both loesses to the Peorian, or in referring the upper buff loess to post-Wisconsin times. However this may be, the Illinoian and Sangamon epochs are post-mid-Pleistocene, from the standpoint of the duration of the Pleistocene, and the fauna represented by the McAdams collection may be regarded as post-mid-Pleistocene.

THE PHYSICAL CHEMISTRY OF THE CRYSTALLIZATION AND MAGMATIC DIFFERENTIATION OF IGNEOUS ROCKS

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III

ON THE QUARTZ, CRYSTALLIZING AT A LATE STAGE, IN QUARTZ-BEARING NORITES, GABBROS, SYENITES, DIORITES, ETC.

In the rocks here mentioned, where the quantity of quartz does not surpass 5 per cent or occasionally—especially in some quartz-diorites[1]—somewhat more, the quartz, as is well known, appears as *Zwischenklemmungsmasse,* or mesostasis, indicating a very late stage of crystallization. If we for convenience' sake only concern ourselves with the gabbroidic rocks, we find this depending on the fact that in a complicated system (Ab+An:ferromagnesian meta-silicates [with other pyroxene components]:Qu), a great deal of plagioclase and pyroxene will crystallize at an early stage if the Qu component is present only in small quantity. By this means the quantity of the Qu component in the mother-liquid increases, and the quartz can only commence forming when a complicated eutectic boundary between Qu and Ab+An and ferromagnesian silicates has been reached.

In this manner the quartz will fill the intervening spaces between the already formed plagioclase and pyroxene individuals. At this late stage of crystallization, however, we have not, as occasionally assumed by some earlier investigators, a crystallization of quartz alone, but, on the contrary, of quartz simultaneously with some plagioclase and ferromagnesian silicate. The fact is that the quartz in the mesostasis often forms a pegmatitic or granophyric intergrowth as well with the plagioclase as with the pyroxene in

[1] Rocks with acid plagioclase, with at least 15 per cent quartz, I do not include in the group of diorites.

question. As an example we refer to Figures 17 and 18 of a quartz-norite from Erteli, Norway, consisting of about 60 per cent labradorite, 35 per cent hypersthene (with a little secondary hornblende), and on the average 5 per cent quartz, but without oxidic iron ore and without biotite. Some parts of the thin section show only $\frac{1}{2}$, 1, or 2 per cent of quartz, but locally, as in the part photographed, the quantity of quartz rises much higher, even to 20 per cent.

The relative proportions of the three minerals of the last stage of crystallization cannot be accurately defined under the microscope. We get the impression, however, that the percentage of

FIG. 17.—Photomicrograph between crossed nicols (24:1).

FIG. 18.—Drawing (27:1)

Hyperitic-structured quartz-norite from Erteli, Norway. Labradorite, with twinning lamellae after the albite law (and quite subordinate after the pericline law). Hypersthene (dark shading) with a little secondary hornblende. A little orthoclase (dotted in diagram) at the periphery of one or two labradorite individuals. The drawing represents about seven-eighths of the photomicrograph.

ferromagnesian silicates in the final product of the solidification is quite small. As an estimate we may rate about 55 per cent labradorite, 35 per cent quartz, and 10 per cent pyroxene.

Their mutual proportions, however, will to a great extent be dependent upon the composition of the plagioclase and the pyroxene.

Originating from the values mentioned, we may imagine a norite or gabbro with 3.5 per cent quartz, where the quartz and

the remaining part of the final product will commence to crystallize only after all in all 90 per cent plagioclase and pyroxene have solidified.

As is well known, a little orthoclase or microcline often appears with the quartz in the mesostasis of the gabbro rocks. The explanation of this fact is quite simple. If the magma contains somewhat more Or ($KAlSi_3O_8$) than is absorbed by the plagioclase, the small surplus of Or will be concentrated in the final magma, and consequently will crystallize together with the quartz and some plagioclase and ferromagnesian silicate in the mesostasis.

In the chapter in Part II on the "Oligoclase-Granite Dikes" of many gabbros and norites, we will show that these dikes represent the final magma, resulting at a very late stage of the crystallization of gabbroidic magmas which contain a little quartz.

ON THE DIFFERENCE BETWEEN THE SOLUBILITY OF THE FERROMAGNESIAN SILICATES IN ACID (GRANITIC) AND BASIC (GABBROIDIC) MAGMAS

We shall commence by mentioning some binary eutectics.

Synthetic determinations, according to the Geophysical Labaratory, Washington:

Qu (crystobalite, *ca.* 1650°):An (1550°)=48 per cent Qu:52 per cent An (1353°).

Qu (crystobalite):Diops (1391.5°)=16 per cent Qu:84 Diops (1362°).

An:Diops=42 per cent An:58 per cent Diops (1270°).

Ab (*ca.* 1100°):Diops=about 97 per cent Ab:3 per cent Diops. (The last statement is according to extrapolation by Bowen.)

Analytical (see above) Qu:Ab=about 28 per cent Qu:72 per cent Ab.

For the Na_2O-rich granites containing pyroxene instead of the usual biotite we must assume that Qu:Ab:Diops (at high pressure) form a ternary eutectic. The same must be assumed also for Qu:An:Diops. The latter system, however, is of subordinate interest petrographically. For the ternary systems Qu:Ab:Diops and Qu:An:Diops, with regard to which we know all the binary eutectics between the individual substances, we shall construct the approximate ternary systems, Figure 19. We here especially fix our attention on the fact that the ternary eutectic Qu:Ab:Diops can contain no more than a few per cent, or probably not even so much, of diopside. On the other hand,

the eutectic boundary-line between diopside (or pyroxene in general) and anorthite, bytownite, and labradorite (see above) contains as much as about 55, 45, and 35 per cent pyroxene respectively, and these figures will only be displaced to a slight degree by the presence of a lesser quantity of quartz, as 2, 5, or 10 per cent.

In magmas of *granitic* composition, with Ab as the plagioclase, diopside must consequently commence crystallizing earlier than plagioclase (albite), even if only as little as a few per cent of diopside are present. In magmas of *gabbroidic* composition (with labra-

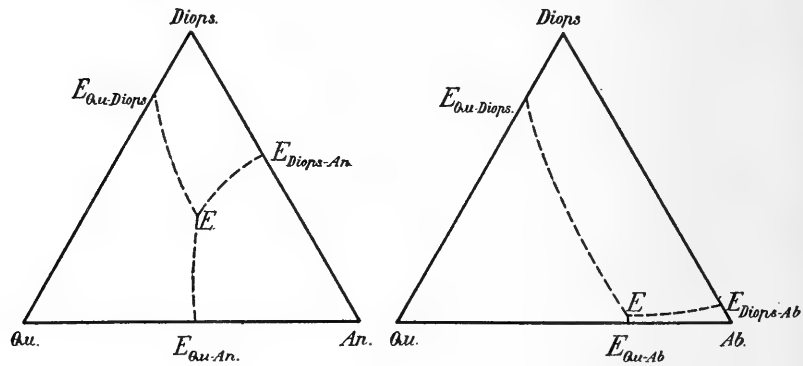


FIG. 19.—Diagrams of the individualization fields Qu:Diops:An, and Qu: Diops:Ab.

dorite or bytownite), on the other hand, plagioclase commences crystallizing earlier than the pyroxene, even if as much as 35–40 per cent of pyroxene is present.

This result with regard to the slight solubility of $CaMgSi_2O_6$ in acid magmas, deduced from physico-chemical foundations, is verified by the petrographical investigation of granitic igneous rocks. And this slight solubility in the acid igneous rocks F̈e applies not only to diopside, but also to the Mg, Ca-, Mg, F̈e, Ca-, or Mg, F̈e-silicates in general.

It is apparent that—

1. The crystallization of the silicates in the granites (with about 70–76 per cent SiO_2) commences with the crystallization of the ferromagnesian silicates (biotite, hornblende, augite, hypersthene) when the latter is present in a quantity of at least a few

per cent. With more basic plagioclase (andesine) somewhat more of the ferromagnesian silicate must be present if the latter is to commence crystallizing earlier than the plagioclase.

2. In the granitic eutectic we only find a trifle MgO, viz., according to the analyses collocated on page 348, we find in magmas with little An only about 0.2 per cent MgO (and in the magmas, somewhat richer in An, possibly as much as about 0.5 per cent MgO).

The very great difference between the degree of solubility of ferromagnesian silicates (mica, pyroxene, or hornblende) in *acid* rocks, consisting chiefly of acid feldspar and quartz, and in *basic* rocks, containing basic plagioclase, is of great petrologic importance.

ORTHORHOMBIC AND MONOCLINIC PYROXENE

In orthorhombic pyroxene we often observe microscopically small lamellae of monoclinic pyroxene, and in monoclinic pyroxene corresponding minute lamellae of orthorhombic. This must (see above, p. 436) be explained as a secondary phenomenon, due to a secretion in the solid phase.

Where *independent* individuals of orthorhombic and monoclinic pyroxene are intergrown, the orthorhombic (hypersthene)—as has often been pointed out by earlier investigators and as I have often observed in norites containing diallage—forms the kernel and the monoclinic (diallage) the surrounding parts. This indicates that the hypersthene was formed earlier than the diallage. Exceptionally also we find large crystals of hypersthene, with quite good idiomorphic outlines, inclosed in the diallage.

We here refer to Figures 20–21 of a norite,[1] containing diallage, from Skjækerdalen in Norway, and consisting of about 5 per cent olivine (according to the determination of the optical character and the axial angle with 20–25 per cent Fe_2SiO_4) in scattered individuals (not represented in the section drawn); about 25 per cent hypersthene (according to optical determination with about 25 per cent $FeSiO_3$); about 25 per cent diallage (optically positive, $2V = ca.\ 65°$, $c:\mathfrak{c} = ca.\ 43°$) with a little primary brown hornblende;

[1] This rock has recently been treated by C. W. Carstens, *Geology of the Trondhjem District*, 1920, p. 101.

about 40 per cent labradorite (about Ab_2An_3); and about 5 per cent pyrrhotite. If we leave out of consideration the olivine, which appears only here and there and whose age we were unable to determine with certainty in the present case, the hypersthene is the oldest mineral, as it shows an idiomorphic contour against all the other minerals—however, with somewhat rounded edges. (See the remarks in a following chapter.)

We especially emphasize that crystals of hypersthene in several places have swum together to small aggregates, showing a synneusis structure (see the left side of Fig. 21), and that the hypersthene shows idiomorphic outlines also against the diallage, with

FIG. 20.—Photomicrograph (15:1)

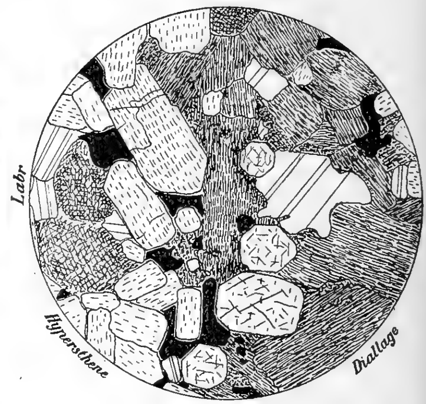


FIG. 21.—Drawing (15:1)

Norite from Skjækerdalen, Norway. Contains hypersthene (in the photograph light gray, in the drawing lightly shaded), diallage (in the photograph darker gray, in the drawing dark shading), and a little brown hornblende (dotted in diagram), labradorite (white in the photograph, showing twin lamellae in the drawing), and some pyrrhotite (black).

which, in the present case, it is not in parallel growth. This cannot be explained otherwise than that the hypersthene had finished forming before the commencement of crystallization of the diallage.

I believe I am right in drawing the conclusion that when hypersthene and diallage appear together in igneous rocks, the hypersthene, regardless of the quantitative proportion between the two minerals, is prevailingly the oldest. Hypersthene (consisting of two chief components) and diallage (consisting of several

components) are consequently related to each other as the components with high and low melting-points in the binary mix-crystal system No. IV (cf. Fig. 36).

PYROXENE AND BIOTITE, RESPECTIVELY HORNBLENDE, IN THE GABBRO ROCKS

We have previously treated one case, viz., the quartz-norite from Romsaas (see pp. 434–35) where *hypersthene* first crystallized, but later stopped forming as the remainder of the ferromagnesian silicate in the magma entered into *biotite*. In another series of norites and gabbros with relatively much orthorhombic or monoclinic pyroxene and usually with only 2–5, seldom up to 8–10 per cent of biotite, we find the latter (see, for example, Fig. 13) partly grown into the *exterior* parts of, and partly grown on to, the pyroxene, indicating that the biotite belongs to a somewhat later stage of crystallization than the pyroxene.

Primary hornblende is lacking in numerous Norwegian norites and gabbros, but appears in others, most often, however, only in a quantity of 5–10 per cent. This primary hornblende, which always, or nearly always, is brown or greenish-brown, and probably throughout contains some titanic acid, often shows a parallel intergrowth with the orthorhombic or monoclinic pyroxene, in such a manner that the hornblende appears at the periphery of the pyroxene. Or the hornblende may have grown as independent individuals on to the pyroxene. This signifies also for the hornblende a stage of forming later than for the pyroxene. The explanation probably is that the original small contents of H_2O in the gabbroidic magma was concentrated in the residual magma by the solidification of a greater or lesser part of the pyroxene, so that biotite, respectively hornblende, was able to individualize. This will be discussed in a later chapter.

OLIVINE, Mg_2SiO_4:Fe_2SiO_4

The melting-point of pure Mg_2SiO_4, according to the determination at the Geophysical Laboratory, Washington, is 1890°. The melting-point of Fe_2SiO_4 lies, according to approximate determination, at *ca.* 1100°. Doelter gives *ca.* 1065°. Mg_2SiO_4 and Fe_2SiO_4 form a continuous mix-crystal series.

a) As there is such an exceedingly great difference between the melting-points of the two components, it must a priori be taken for granted that the binary system $Mg_2SiO_4:Fe_2SiO_4$ belongs to type I (without a maximum or minimum).

b) In zonal olivine we find, according to a number of investigators (Sigmund, Becke, Stark),[1] Mg_2SiO_4 concentrated in the

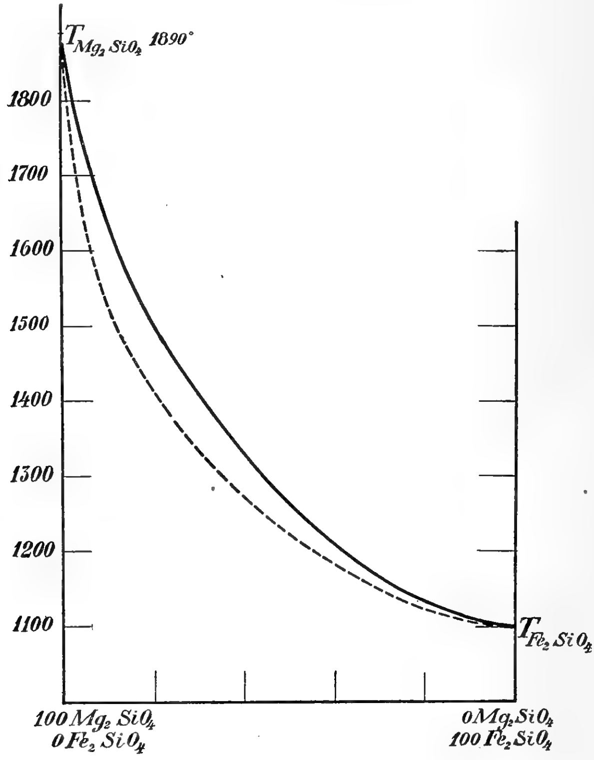


Fig. 22.—Schematic melting-diagram $Mg_2SiO_4:Fe_2SiO_4$ (percentages by weight)

kernel, and Fe_2SiO_4 in the peripheral zone. Stark found in a zonal olivine from basalt containing 19.5 per cent Fe_2SiO_4 in the kernel and 34.5 per cent in the exterior zone, consequently a marked difference, but not by far so great a difference as between Ab:An in the plagioclase of corresponding rocks. These investi-

[1] See *Tscherm. min. u. petr. Mitt.*, Vols. XVI, XVII, and XVIII.

gations of zonal olivine apply to the common rock-forming olivine, with at most 35–40 per cent Fe_2SiO_4 (and rest Mg_2SiO_4).

c) The earlier approximate determinations, especially by Doelter, show for olivine with increasing percentages of Fe_2SiO_4, continuously decreasing "melting-points" (ↄ:melting-point interval).

d) The later investigations prove *inter alia* that even so small a percentage of Fe_2SiO_4 as 10–15 per cent lowers the melting-point interval of the olivine considerably below the melting-point of Mg_2SiO_4. A maximum of the melting-curve, which in this case must have occurred at predominant Mg_2SiO_4 and little Fe_2SiO_4, is therefore out of the question.

In conformity with all the above observations I believe myself justified in illustrating the binary system Mg_2SiO_4:Fe_2SiO_4 by the foregoing sketch (Fig. 22); it must be emphasized, however, that the course of the curve is only sketched.[1] As the zonal structure of the olivine is far less prominent and there is less difference between the two components than in the plagioclase in rocks with about the same cooling-rate, the difference between the liquidus and solidus curves must be less for Mg_2SiO_4:Fe_2SiO_4 than for An:Ab.

OLIVINE AND PLAGIOCLASE

As explained in my earlier treatise "Die Silikatschmelzlösungen," I and II (1903–4), olivine and anorthite do not crystallize at the pressure of one atmosphere in molten masses of certain intermediate mixtures of Mg_2SiO_4 and $CaAl_2Si_2O_6$, for here new minerals are formed, chiefly melilite and spinel. In conformity with this, in the treatise above cited, I discussed for basic silicate melts with anorthite and olivine as the two extremes, not the individualization boundary between olivine and anorthite, but between (*a*) olivine and melilite, and (*b*) melilite and anorthite.

O. Andersen (at that time with the Geophysical Laboratory, Washington), in his treatise "The System Anorthite:Forsterite:Silica,"[2] arrived at the same results, for by intermediate mixtures of An and Mg_2SiO_4—viz., between 90 An:10 Mg_2SiO_4 and 54 An:46 Mg_2SiO_4—he obtained crystallized spinel. At *high pressure,*

[1] A minimum in the neighborhood of Fe_2SiO_4 is not excluded but very improbable.

[2] *Loc. cit.*

as in the crystallization of deep-seated igneous rocks, we find a different case. *Forellengestein* or troktolite, for instance, consists of very basic plagioclase (anorthite-bytownite) and olivine. Some anorthosites show predominant plagioclase (bytownite, labradorite) and in addition some olivine, and some olivine gabbros consist chiefly of basic plagioclase and olivine with only a small amount of monoclinic or orthorhombic pyroxene. On remelting several of these rocks at atmospheric pressure, more or less melilite results, and in addition frequently some spinel, and occasionally also other minerals. At the pressure of one atmosphere, consequently, quite a different mineral combination results from the magmas here mentioned than the combination occurring in the deep-seated rocks. We shall in a following chapter discuss the cause of this. Here we shall only fix our attention on the fact that the melting of intermediate mixtures of $CaAl_2Si_2O_8$ and Mg_2SiO_4 at the pressure of one atmosphere does not give us the required information of the individualization boundary between plagioclase (anorthite) and olivine, which takes place at high pressure. In the question in hand, therefore, we must use the analytical method.

Petrographical experience shows that the olivine in most of the igneous rocks crystallized at a very early stage. Because of this fact, many petrographers have the conception that olivine takes an exceptional position with regard to the sequence of crystallization and that this mineral always crystallizes as No. I of the silicate minerals. This is, however, a misconception.[1] The reason why the olivine in numerous cases commences crystallizing earlier than the other silicates is that the individualization boundary between olivine and plagioclase or pyroxene, or between olivine and plagioclase plus olivine, lies at the point of relatively little olivine, and that most rocks contain more olivine than the individualization boundary mentioned. But if less olivine is present the solidification commences with the crystallization of one of the other minerals. Returning to the rocks which consist only or nearly only of olivine

[1] As early as in my first mineral-synthetic work ("Studies on Slags," 1885) I opposed this misconception, which *inter alia* also has been expressed in petrographical works of later years. But when a misconception of an authoritative character has been impressed on one's consciousness, it may require a decade of years to uproot it.

and (basic) plagioclase, it has already been indicated by A. Harker[1] that of these two minerals, the one present in quantity above a certain limit first commences crystallizing. This is also in accordance with my own investigations.

I find it superfluous to discuss the plagioclase rocks with considerable olivine, and as a consequence with crystallization of an

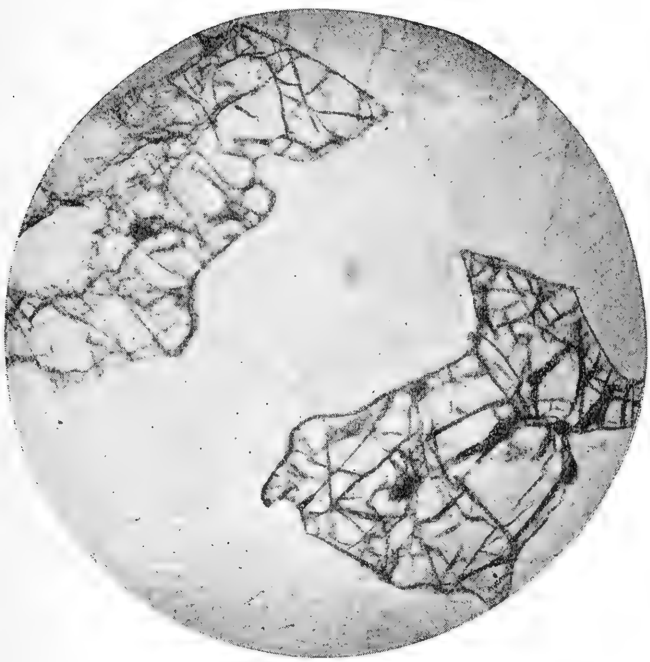


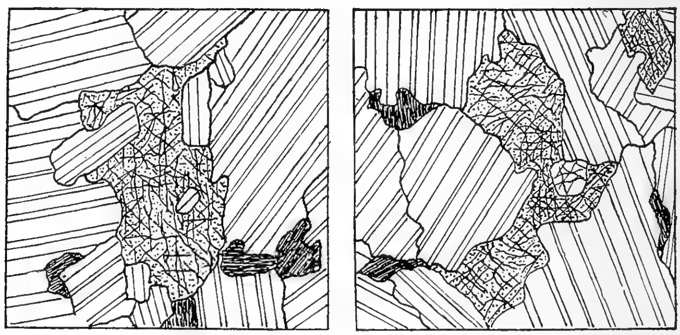
Fig. 23.—Photomicrograph (20:1)

essential part of the olivine before the commencement of the solidification of the plagioclase. But I am going to discuss the inverse proportion, much plagioclase and little olivine, and choose as an example an olivine-bearing labradorite rock from the Ekersund field. This consists of nearly 90 per cent labradorite (Ab_1An_1), about 7 per cent olivine (optically negative; $2V = ca.\ 85°$, consequently with about $0.25\ Fe_2SiO_4 : 0.75\ Mg_2SiO_4$), besides a little diallage. (See photograph Fig. 23 and drawing Fig. 24.)

[1] *The Natural History of Igneous Rocks*, 1909, p. 170.

The olivine here shows no sign of idiomorphism, but this is often the case with labradorite. Laths (parallel to 010) of the latter here and there (Fig. 24*a*) protrude into the olivine, and the straight-edged boundary which we often observe (Fig. 23) between the labradorite and the olivine is not caused by the crystal-limit of the olivine, but by that of the plagioclase. The feldspar must thus here have commenced crystallizing at an earlier stage than the olivine, which, in the same manner as the diallage in Figures 24*a* and 24*b*, and as the hypersthene in Figure 14, chiefly forms an intervening mass between relatively large labradorite individuals.

In the *olivine-rich olivine hyperites* (hyperitic-structured olivine gabbros) with about 25–30 per cent olivine, 60 per cent



a *b*

FIGS. 24*a* and 24*b*.—Drawings (20:1)

Anorthosite from Ekersund, Norway, containing *ca.* 90 per cent labradorite (Ab_1An_1), *ca.* 7 per cent olivine (dotted in Fig. 24), and a little diallage (dark shading in Fig. 24, not seen in the photograph). The straight lines in Figure 23 represent the idiomorphic contours of the labradorite against the olivine.

labradorite, 10–20 per cent diallage, and a little magnetite, etc., the olivine chiefly appears in synneutic individuals with very good idiomorphism against the diallage and partial idiomorphism also against the plagioclase (see, for instance, Fig. 33). Considering only the silicate minerals, we find consequently that first a good deal of olivine solidified, then the labradorite, and at a later stage the diallage also commenced crystallizing.

In the *olivine-poor olivine hyperites*, with only 5–10 per cent olivine, we find, on the other hand, in the relation between the olivine and the labradorite quite a different structural phenome-

non, viz., lath-shaped labradorite individuals protruding into the olivine, and the latter shows no sign of idiomorphism even against the labradorite. The olivine chiefly fills the intervening spaces between the plagioclase laths in the same manner as in the anorthosite illustrated in Figures 23 and 24. We refer, for instance, to Figures 48 and 49 where the original structure, however, is partly effaced by the here quite strongly developed reaction rims between the olivine and the labradorite. In these olivine-poor olivine hyperites a greater or lesser part of the plagioclase must according to the structure have crystallized earlier than the olivine.

On the basis of the crystallization sequence we draw the conclusion that the individualization boundary at high pressure between the olivine (about $1Fe_2SiO_4 . 2Mg_2SiO_4$) and basic plagioclase lies at little olivine and much plagioclase. As an estimate we may set the individualization boundary (by weight) at 0.15 Oliv.:0.85 Ab_1An_1, 0.25 Oliv.:0.75 Ab_1An_3, and probably about 0.35 Oliv.:0.65 An. That these statements are approximately right is verified by a study of the "orbicular gabbro" from Debesa, California, described by A. C. Lawson,[1] which carries orbs with changing shells (challotes) of radially arranged olivine together with bytownite. This structure must depend on a *simultaneous* crystallization of the two minerals, that is to say, a crystallization along a eutectic boundary-line. From the quantitative analysis we calculate the composition of this boundary at about 0.34 Oliv. (0.35 Fe_2SiO_4. 0.65 Mg_2SiO_4):0.66 Plag. ($Ab_{16}An_{84}$, consequently about Ab_1An_5.)

The quartary system Ab:An:Mg_2SiO_4:Fe_2SiO_4 separates at high pressure into two individualization fields, viz., Ab+An and $Mg_2SiO_4+Fe_2SiO_4$. We may here apply the same general considerations as for the system Ab+An and $CaMgSi_2O_6+CaFeSi_2O_6$ (see p. 442).

OLIVINE AND PYROXENE

Olivine and monoclinic pyroxene.—Forsterite, Mg_2SiO_4 (melting-point=1890°) and diopside, $CaMgSi_2O_6$ (melting-point=1391°), form, according to N. L. Bowen,[2] a eutectic, 88 per cent diopside: 12 per cent forsterite with melting-point=1386°, consequently

[1] *Univ. of Cal. Publications, Dept. of Geol.*, Vol. III (1904).

[2] "The Ternary System Diopside-Forsterite-Silica," *Amer. Jour. of Sci.*, Vol. XXXVIII (1914).

only 4–5° below the melting-point of diopside. According to my earlier studies on slags, the individualization boundary between augite ($CaMgSi_2O_6$ with some $CaFeSi_2O_6$, $MgAl_2SiO_6$, etc.) and olivine (Mg_2SiO_4 with some Fe_2SiO_4 and Mn_2SiO_4, consequently with a much lower melting-point than pure Mg_2SiO_4) lies at about 70 per cent of augite: 30 per cent of olivine. At this individualization boundary a decrease of the melting-point appears according to experimental investigations.[1] We accordingly here have to deal with a eutectic boundary-line between the two mix-crystals. The location of this boundary is chiefly dependent upon the composition of the olivine, and this is probably due to the fact that the melting-point interval of the olivine is considerably lowered by some Fe_2SiO_4. (See Fig. 22.)

Regarding $Mg_2SiO_4 : MgSiO_3$, and especially with regard to the dissociation of $MgSiO_3$ at the pressure of one atmosphere and at high temperature (1577–1555°), some forsterite first being solidified from melted $MgSiO_3$, we refer to Bowen's investigations which are discussed in a following chapter. I here arrive at the result that the dissociation of $MgSiO_3$, determined at the pressure of one atmosphere, cannot be transferred to apply to (Mg, *Fe*) SiO_3 or (Mg, Fe) SiO_3 at *high* pressure.

In igneous rocks consisting of olivine and orthorhombic or monoclinic pyroxene, the olivine chiefly appears in idiomorphic individuals when the olivine forms at least one-third, and the pyroxene at most two-thirds. This fact is so well known that I find it superfluous to give special examples. But when the olivine is present only in small quantity, the sequence of crystallization is quite turned about. Such rocks, with predominant pyroxene and quite little olivine, are in themselves rare, and furthermore the olivine in these rare rocks is only exceptionally fresh enough to permit a detailed study of the original structure of the rock. As far as I know, these rocks have not previously attracted any special attention; I shall therefore discuss a couple of examples.

In a hypersthenite-norite from Nonaas-Litland in Hosanger, Norway (see analysis in the chapter on norites in Part II), a con-

[1] See "Silikatschmelzlös.," II.

cretion of an *olivine-bearing hypersthenite* (or olivine-augite-hornblende-hypersthenite) appears locally (at Nonaas), with a chemical composition:

about	47–48	per cent SiO_2
about	0.5–1	per cent TiO_2
about	5–6	per cent Al_2O_3
about	2.5–3	per cent Fe_2O_3
about	12–13	per cent FeO
about	17–18	per cent MgO
about	11–12	per cent CaO
about	0.5–1	per cent alkali

This rock, which is quite fresh, consists mineralogically of about 75 per cent pyroxene, viz., about 50 per cent hypersthene (optically negative, $2V = ca.\ 80°$, that is to say, with stoechiometric 25–30, closest at about 27 per cent $FeSiO_3$) and about 25 per cent monoclinic pyroxene (extinction-angle 39–40°); on an average about 12 per cent olivine (optically negative, $2V = ca.\ 83°$, consequently with stoechiometric[1] about 30 per cent Fe_2SiO_4); about 12 per cent intensively greenish-brown, primary hornblende; and in addition locally 0.5 per cent biotite and about 0.5 per cent plagioclase, the latter only here and there as mesostasis of quite subordinate importance. Oxidic iron ore is entirely lacking. Apatite and pyrites appear in small quantity. On the photomicrograph, Figure 25, the light-streaked mineral chiefly consists of hypersthene and a few individuals of monoclinic pyroxene. Uppermost to the right we see a hypersthene individual, cut at right angle to the *c*-axis.

The mineral which on account of the strong absorption of light shows dark-gray in the photograph is hornblende, and the mineral which appears white is olivine, quite fresh, without any sign of serpentinization or other change. The pyroxene individuals show, as well in length as in cross-section, quite good idiomorphism against the olivine. The same, though not quite so distinctly, is the case with the hornblende, which belongs to a somewhat later stage of crystallization than the hypersthene (see above). On the other hand, the olivine does not show even a sign of idiomorphism against the hypersthene and the other minerals, but forms—in

[1] The optical determinations have been undertaken by Docent Carstens and myself in collaboration.

the same manner as the plagioclase in the gabbro rocks which are especially rich in hypersthene or diallage—a mass of putty or a *Zwischenklemmungsmasse* (mesostasis) between the ferromagnesian silicate minerals.

So we arrive at the result that in the rock in question, with little olivine and predominant hypersthene, with monoclinic pyroxene and some hornblende, the olivine belongs to the *last* stage of crystallization.

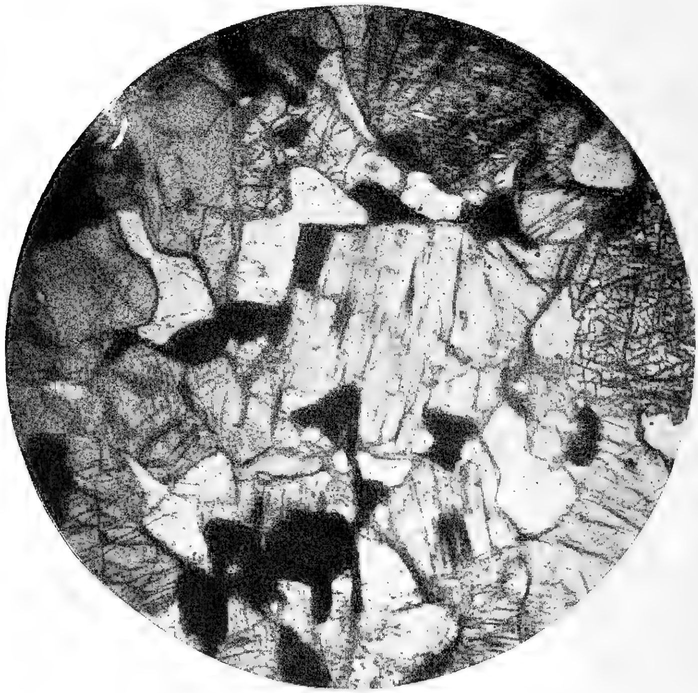


FIG. 25.—Photomicrograph (24:1) of olivine-bearing hypersthenite with some diallage and hornblende, from Nonaas, Hosanger, Norway. The light-gray mineral is hypersthene with some diallage, the dark-gray is hornblende, and the white olivine.

Also in some other olivine-bearing pyroxenites, moderately rich in iron, with predominant hypersthene, or diallage or both, and in addition with only a quite small amount, say 10 or 15 per cent, of olivine, the structure proves the late commencement of the crystallization of the olivine.

If we pass on to the dunites and saxonites *poor in iron*, usually with 40–43 per cent SiO_2, 42–48 MgO, 7–10 FeO, 0–1.5 CaO, and

a little Al_2O_3, we find that the dunites consist nearly exclusively of olivine, and the saxonites usually of predominant olivine with only quite little bronzite or enstatite-bronzite.

Olivine-bronzite (or enstatite) rocks with predominant orthopyroxene and quite little olivine are very rare, and personally I have only once (1893) found such a rock, viz., as a local facies of a peridotite at Esjeholmen near Næsö, in the Hestmandö district, near the Polar Circle in the northern part of Norway.[1]

As illustrated in Figure 26, there here appear rosettes of radially arranged bars of enstatite which in most places are entirely unchanged, but in some places somewhat altered to tremolite, clinochlorite, talc, and magnesite.

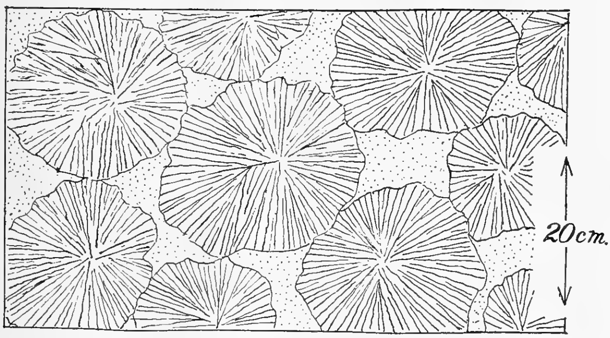


FIG. 26.—Enstatite-olivine rock from Esjeholmen, near Næsö, Hestmandö district, northern Norway. Contains great rosettes of enstatite and an intervening mass of olivine with some enstatite. (1/10 nat. size.)

The bars of enstatite in these rosettes may reach a length of 1 dm., or somewhat more, and the enstatite rosettes may have the size of the head of a full-grown man. The intervening mass between the enstatite rosettes consists of olivine and enstatite. In the entire rock we may reckon about 80 per cent enstatite rosettes and only about 20 per cent intervening mass, consequently for the whole rock about 90 per cent enstatite and 10 per cent olivine, in addition to a minimal quantity of chromite. I cannot explain this structure otherwise than that at first the enstatite of the large rosettes was formed, and later the intervening mass, consisting of olivine and enstatite. The result of this is that the sequence of

[1] See a treatise by myself in *Zeitschr. f. prakt. Geol.*, 1894, pp. 389–92, and a treatise by C. W. Carstens, "Norske peridotiter," I, *Norsk geol. tidskr.*, Vol. V (1918).

crystallization in the deep-seated rocks of olivine and monoclinic, or orthorhombic, pyroxene must be explained by the same laws as for the other ordinary rock-forming minerals, and that the individualization boundary lies at predominant pyroxene: little olivine.

ON THE RELATION BETWEEN MgO AND FeO (OR BETWEEN Mg-SILICATE AND Fe-SILICATE) IN THE FERROMAGNESIAN SILICATES, CRYSTALLIZED FROM THE SAME MAGMA

We include a selection of analyses of the primary ferromagnesian minerals appearing in the same rock:

	No.	SiO_2	TiO_2	Cr_2O_3	Al_2O_3	Fe_2O_3	FeO	MnO	MgO	CaO	Na_2O	K_2O	P_2O_5	H^2O	Total
Websterite (No. 49) with Bronzite (*b*) and Diopside (*c*)															
Websterite.....	49	52.55	0.14	0.44	2.71	1.27	4.90	0.24	20.39	16.52	0.27		Tr.	1.09	100.52
Bronzite.......	*b*	54.53	.....	0.30	1.93	1.70	8.92	0.28	29.51	2.25	.....	....	.....	1.14	100.56
Diopside.......	*c*	51.80	0.13	0.51	2.21	1.29	3.50	Tr.	17.76	20.99	.....	....	.....	0.65	98.84
Pikrite (No. 50) with Bronzite (*b*) and Diopside (*c*)															
Pikrite.........	50	37.12	0.40	......	4.96	8.92	7.62	0.40	26.92	6.14	0.40	0.49	0.10	5.04	98.60
Bronzite.......	*b*	54.20	.....	0.29	2.05	......	10.08	.....	29.00	2.49	.....	....	.....	0.42	98.53
Diopside.......	*c*	52.63	.....	0.72	2.24	......	6.84	.....	17.30	20.04	.....	....	.....	0.57	100.34
Hypersthene Diabase (No. 51) with Hypersthene (*b*) and Augite (*c*)															
Diabase........	51	50.86	.....	......	7.95	......	7.41	.....	18.93	13.29	1.34	0.20	.....	.....	100.00
Hypersthene...	*b*	52.16	.....	......	3.00	0.43	15.16	0.36	21.89	5.94	0.16	0.04	.....	0.08	99.24
Augite.........	*c*	49.33	.....	......	9.15	0.27	9.05	.....	14.58	16.36	0.55	0.19	.....	0.25	99.73
Diallage-Hornblende-Gabbro (No. 52) with Diallage (*e*) and Brown, Primary Hornblende (*d*)															
Gabbro.........	52	53.19	0.25	......	11.42	1.74	8.15	0.17	10.84	11.01	2 97	0.31	.....	.....	100.05
Diallage.......	*e*	51.23	0.86	......	2.75	3.88	14.09	Tr.	17.05	9.74	.....	....	.....	1.75	101.35
Hornblende....	*d*	48.04	.....	......	8.98		16.44	.....	18.87	6.72	.....	....	.....	.....	99.06
Dacite (No. 53) with Hypersthene (*b*), Biotite (*e*), and Ilmenite (*f*)															
Dacite.........	53	63.27	1.30	......	16.50	0.68	5.10	0.03	2.48	4.18	2.36	2.68	0.15	0.61	99.50
Hypersthene...	*b*	50.42	3.51	......	4.06	2.10	23.54	0.24	13.04	1.30	Tr.	0.69	0.92	0.16	99.98
Biotite.........	*e*	39.86	7.95	......	11.13	1.39	18.10	0.58	9.88	Tr.	0.35	6.73	.....	3.63	99.60
Ilmenite.......	*f*	.....	67.28	......	.....	......	31.92	.....	0.80	.....	.....	....	.....	.....	100.00
Peridotite (No. 54, Kimberlite, Strongly Decomposed) with Olivine (*g*), Garnet (*h*), and Ilmenite (*f*)															
Peridotite......	54	29.81	2.20	0.43	2.01	5.16	4.35	0.23	32.41	7.69	0.11	0.20	0.35	8.92	100.86
Olivine........	*g*	40.04	0.07	0.24	0.39	2.36	7.14	0.20	46.68	1.16	0.08	0.21	0.04	0.80	99.42
Garnet........	*h*	41.32	0.16	0.91	21.21	4.21	7.93	0.34	19.32	4.94	0.07	....	.....	0.17	100.58
Ilmenite.......	*f*	0.76	49.32	0.74	2.84	9.13	27.81	0.20	8.68	0.23	0.19	....	.....	0.20	100.00

No. 49: Websterite, composed only of bronzite and diallage. Hebbville near Baltimore. G. H. Williams, *Amer. Geologist,* Vol. VI; F. W. Clarke, *U.S. Geol. Sur. Bull. 228* (1904), p. 51.—No. 50: Pikrite. Schwarzenstein, Fichtelgebirge. In the rock 0.09 per cent CO_2. Gümbel, *Geogr. Beschreibung Bayerns,* 1879. (Rosenbusch, *Gesteinslehre,* p. 352.)—No. 51: Hypersthene diabase. Twins by Rapidan, Virginia. Campbell and Brown, *Bull. Geol. Soc. Am.,* II.—No. 52: Diallage-hornblende-gabbro. Veltein, in the Alps. Küchler, *Chemie der Erde,* I, 1914. The analysis of the rock from Hecker, *Neues Jahrb. f. Min., Geol. u. Pal.,* Beil., Bd. XVII (1903).—No. 53: Hypersthene-biotite-dacite. Upway, Victoria. Skeats, *Quart. Jour. Geol. Soc. London,* 1910. In the rock 0.16 per cent S. In the biotite H_2O+, 3.20 and $H_2O\div$, 0.43 per cent.—No. 54: Kimberlite, strongly decomposed. From Elliot County, Kentucky. In the rock 8.92 per cent H_2O, 6.66 CO_2, 0.28 SO_3, 0.05 NiO. Diller, *U.S. Geol. Soc. Bull. 38,* and *Amer. Jour. Sci.,* 3d Series, Vol. XXXII; Clarke, *U.S. Geol. Soc. Bull. 228,* p. 66.

OLIVINE AND ORTHORHOMBIC PYROXENE

We shall commence with some analyses of olivine and bronzite from olivine nodules in basalts. We shall base the calculations for the bronzite, often somewhat decomposed, on the entire quantity of iron found analytically. In reality a little, but only very little, iron in the bronzite will appear as Fe_2O_3.

The stoechiometric relation between MgO and FeO in olivine and bronzite for olivine nodules:

Styria (Kappenstein, etc.)	55*a*	Olivine	1 MgO:0.11 FeO
	55*b*	Bronzite	1 MgO:0.14 FeO
	56*a*	Olivine	1 MgO:0.11 FeO
	56*b*	Bronzite	1 MgO:0.10 FeO
Dreizer Weiher, Eifel	57*a*	Olivine	1 MgO:0.09 FeO
	57*b*	Bronzite	1 MgO:0.11 FeO
	58*a*	Olivine	1 MgO:0.11 FeO
	58*b*	Bronzite	1 MgO:0.10 FeO
Stempel, Marburg	59*a*	Olivine	1 MgO:0.10 FeO
	59*b*	Bronzite	1 MgO:0.11 FeO
Kaiserstuhl, Baden	60*a*	Olivine	1 MgO:0.09 FeO
	60*b*	Bronzite	1 MgO:0.10 FeO
Reihenweiler, Alsace	61*a*	Olivine	1 MgO:0.14 FeO
	61*b*	Bronzite	1 MgO:0.14 FeO

Nos. 55*a, b,* and *c:* Kukurzenkezel near Kappenstein: Schadler, *Tscherm. Mitt.,* Vol. XXXII (1914).—No. 56*a* and *b:* Schiller, Becke, *Tscherm. Mitt.,* Vol. XXIV.—Nos. 57*a* and *b:* Th. Kierulf, *Bischof's chem. Geol.,* and *Pogg.*

Ann., Vol. CXLI.—Nos. 58*a*, *b*, and *c:* Philipp, *Neues Jahrb. f. Min.*, etc., 1871, and Rammelsberg, *Pogg. Ann.*, Vol. CXLI.—Nos. 59*a*, *b*, and *c:* Bauer, *Neues Jahrb. f. Min.*, etc., 1891, II.—Nos. 60*a* and *b:* Knop, *Neues Jahrb. f. Min.*, etc., 1877.—Nos. 61*a* and *b:* Linck, *Zeitschr. f. Kryst. u. Min.*, Vol. XVIII.

The value of FeO in the bronzite should be reduced a little throughout, probably about one-tenth in most cases.

Joh. Schiller has discussed the question in hand in a special treatise,[1] partly on the basis of several of the analyses here given of olivine nodules in basalts, and partly on the determination of the chemical composition of the two minerals on the basis of the axial angle and optical character. He comes to the result that MgO and FeO in the feldspar-free rocks are quite evenly distributed in the olivine and orthorhombic pyroxene, and this conclusion is confirmed by my own investigations. But with regard to the rocks containing feldspar he supposes a relative, sometimes even a relatively extensive, enrichment of MgO in the olivine. The observations on which he bases this last construction, however, are few, and in my opinion rather dubious.[2]

Olivine and orthopyroxene,[3] isolated from a series of peridotites poor in iron (saxonites, olivine-schists, etc.) with only very little Al_2O_3, Fe_2O_3, and CaO, show:

Olivine, 1 MgO:0.08, 0.08, 0.08, 0.10, 0.11 FeO;
Orthopyroxene, 1 MgO:0.07, 0.07, 0.07, 0.10, 0.12, 0.12, FeO.

In peridotites, a little richer in iron, and at the same time carrying somewhat more Al_2O_3, Fe_2O_3, and CaO, we find:

Olivine, 1 MgO:0.15, 0.21 FeO;
Orthopyroxene, 1 MgO:0.13, 0.15, 0.16 FeO.

As well with regard to the short report above as to Schiller's investigations, MgO and FeO in the feldspar-free rocks in question are quite evenly divided between the two minerals. The various lesser differences—which would indicate a small relative enrichment

[1] *Tscherm. Mitt.*, Vol. XXIV (1905).

[2] Especially for the extremely low FeO-contents in olivine from an olivine-gabbro from Tilaï, Ural.

[3] Enstatite-bronzite-hypersthene deserves, in the same manner as anorthite-bytownite-labradorite-andesine-oligoclase-albite, a common term, and as such I will use "orthopyroxene," that is to say, pyroxene belonging to the orthorhombic system. I believe I occasionally have heard or read this term before, so the proposition is not originally mine.

of MgO in the olivine in some cases, and in the orthopyroxene in others—approximately balance, and probably depend chiefly on the source of errors connected with the determinations.

If we pass on to the *gabbroidic* rocks, we find that hypersthene, in the common anchi-eutectic norites, usually shows—as well on the basis of the analysis of isolated hypersthene, as by the determinations of the axial angle undertaken by earlier investigators and by myself—a composition between about 30 and 38–40 per cent $FeSiO_3$ (stoechiometric). And the olivine shows, as well on the basis of the analyses of isolated material, as on my own determinations of the axial angle, about 32–35 per cent Fe_2SiO_4.

In the hypersthenite-norites (with only relatively little plagioclase) we usually find, however, a relatively lower percentage of iron, as well in the rock as in the separated silicate minerals. This is discussed more elaborately in Part II. We shall include a couple of separate determinations: The thin secretion of a hypersthenite-norite, above mentioned (Fig. 25), consisting chiefly of hypersthene, augite, hornblende, and olivine shows:

Hypersthene, optically negative, $2V = ca.$ 80°, gives 25–30 (about 27) per cent $FeSiO_3$;

Olivine, optically negative, $2V = ca.$ 83°, gives about 30 per cent Fe_2SiO_4.

Olivine-carrying norite with only about 40 per cent labradorite from Skjækerdalen (Figs. 20–21):

Hypersthene, optically negative, $2V = ca.$ 80–85°, gives about 25 per cent $FeSiO_3$;

Olivine, optically negative, $2V = ca.$ 85–88°, gives 20–25 per cent Fe_2SiO_4.

Also in the igneous rocks containing feldspar, we find approximately the same MgO:FeO proportion in both minerals. Any relative enrichment of MgO in the olivine is usually not to be found.

ORTHORHOMBIC AND MONOCLINIC PYROXENE

As special Fe_2O_3 determinations are lacking in several cases and in others are little instructive on account of a later oxidation, we in both minerals originate from the entire percentage of iron, this giving a quite true image of the relative proportions of MgO and FeO in the two minerals. On account of the small percentage of Fe_2O_3 the statements for FeO ought, however, for the

orthorhombic pyroxene to be reduced about one-tenth, and for the monoclinic pyroxene, which throughout contains a little more Fe_2O_3, about one-eighth. This correction, however, is of small extent.

The stoechiometric proportion between MgO and FeO in orthorhombic and monoclinic pyroxene from the same rock:

Olivine nodules from basalts	Marburg	59*b*	Bronzite	1 MgO:0.11 FeO
		59*c*	Diopside	1 MgO:0.07 FeO
	Dreizer Weiher	58*b*	Bronzite	1 MgO:0.11 FeO
		58*c*	Diopside	1 MgO:0.14 FeO
	Kaiserstuhl	60*b*	Bronzite	1 MgO:0.10 FeO
		60*c*	Diopside	1 MgO:0.16 FeO
	Kappenstein	55*b*	Bronzite	1 MgO:0.14 FeO
		55*c*	Diopside	1 MgO:0.18 FeO
Websterite		49*b*	Bronzite	1 MgO:0.20 FeO
		49*c*	Diopside	1 MgO:0.15 FeO
Pikrite		50*b*	Bronzite	1 MgO:0.20 FeO
		50*c*	Diopside	1 MgO:0.22 FeO
Hypersthene Diabase		51*b*	Hypersthene	1 MgO:0.40 FeO
		51*c*	Augite	1 MgO:0.35 FeO

If we deduct the small, analytically found figures for Fe_2O_3 from the last rock (No. 51), we have:

Hypersthene, No. 51*b*, 1 MgO : 0.39 FeO;
Augite, No. 51*c*, 1 MgO : 0.35 FeO.

The seven double analyses of orthorhombic and monoclinic pyroxene give approximately the same proportions between MgO and FeO, in some cases a little difference in one, and in some in the other direction, but there is no particularly constant enrichment of one component in either of the two minerals. A series of analyses shows that where orthorhombic and monoclinic pyroxene appear as primary formations in the same rock, the monoclinic is characterized by a somewhat higher percentage of TiO_2, Cr_2O_3, Al_2O_3—and probably also of Fe_2O_3—than the orthorhombic.

DIALLAGE AND PRIMARY BROWN HORNBLENDE

Küchler's two analyses from a diallage-hornblende gabbro (No. 52) show, the total quantity of iron being reckoned as FeO:

Diallage, No. 52*c*, 1 MgO : 0.57 FeO;
Hornblende, No. 52*d*, 1 MgO : 0.48 FeO.

If we deduct 3.88 per cent Fe_2O_3 in the diallage and, at an estimate, 2.0 per cent Fe_2O_3 in the hornblende, we get:

Diallage, No. 52*c*, 1 MgO : 0.46 FeO;
Hornblende, No. 52*d*, 1 MgO : 0.42 FeO,

consequently, as emphasized by Küchler, about the same MgO: FeO proportion in both minerals.

HYPERSTHENE AND BIOTITE

The two analyses from a dacite (No. 53) show:

Hypersthene, No. 53*b*, 1 MgO : 1.00 FeO;
Biotite, No. 53*e*, 1 MgO : 1.02 FeO,

consequently exactly the same MgO:FeO proportion in both minerals.

If for the Romsaas quartz-orbicular-norite, which in the entire rock only contains about 0.5 per cent Fe_2O_3, we assume as an estimate 1 per cent Fe_2O_3 in the biotite, we get:

Hypersthene, No. 41, 1 MgO : 0.39 FeO;
Biotite, No. 42, 1 MgO : *0.27 FeO.*

Even if the last figure is not quite exact, relatively somewhat less FeO appears in the biotite than in the hypersthene.

When there is a simultaneous appearance of biotite and hypersthene in the same rock, the biotite seems throughout to carry considerably more TiO_2 than the hypersthene.

The summary, here briefly stated, verifies the earlier conclusion, especially by A. Merian[1] (1884), W. Wahl[2] (1906), and Küchler (*loc. cit.*, 1914), viz., that the composition of the ferromagnesian silicates depends quite simply upon the composition of the entire rock or magma, and further that the relations between MgO and FeO (or Mg-and Fe-silicate) in two from the same magma crystallizing ferromagnesian silicates such as olivine:orthorhombic

[1] "Analysen gesteinsbildender Pyroxene," *Neues Jahrb. f. Min.*, etc., Beil., Bd. III (1884).

[2] *Die Enstatitaugite* (dissertation), Helsingfors, 1906; *Tscherm. Mitt.*, Vol. XXVI (1907).

pyroxene, orthorhombic:monoclinic pyroxene, diallage:primary hornblende, hypersthene:biotite, are not subject to extensive variations. We may find a little variation sometimes in one and sometimes in the other direction, but this may be due in part to inaccurate determinations. But all in all, we here have approximately the same MgO:FeO proportions in both minerals. We especially emphasize that no mineral is characterized by a constant relative enrichment either of MgO or FeO. Lesser variations, with regard to the MgO:FeO proportion, by two or still more ferromagnesian silicates, crystallizing from the same magma, may be due to a series of factors, of which we may mention the horizontal distance between the liquidus and solidus curves (or the difference between the a:b proportion in the first crystallized mix-crystal and in the liquid phase); the degree of equilibrium between the solid and liquid phases; the electrolytic dissociation.

A *small* horizontal difference between the liquidus and solidus curves, and a nearly complete equilibrium between the solid and liquid phases will cause nearly the same MgO:FeO proportion between the segregated ferromagnesian silicates and the magma, and consequently also between the ferromagnesian silicates mutually.

As well in olivine as in orthopyroxene and diopside-hedenbergite, the Mg-silicate is concentrated in the first mix-crystal. By more or less incomplete equilibrium between the liquid and solid phases—as in the dike and effusive rocks—we may expect a relative enrichment of MgO in the mineral which first commenced crystallizing. With two ferromagnesian silicates we may generally expect a more evenly distributed MgO : FeO proportion among deep-seated rocks with complete or nearly complete equilibrium between liquid and solid phases than among dike and effusive rocks.

Addition.—Also in *ilmenite* a little MgO enters, viz., as $MgTiO_3$. In this manner ilmenite No. 54*f* from a peridotite (with about 0.12 FeO:0.88 MgO in the entire rock) shows not less than 8.68 per cent MgO, or 0.64 FeO:0.36 MgO. The ilmenite from the labradorite rock near Ekersund (with about 0.5 FeO:0.5 MgO

in the entire rock) usually contains 3 to 4, up to 5.14 per cent MgO (the last analysis equivalent to 0.78 FeO:0.22 MgO). In the dacite No. 53, with 0.53 FeO:0.47 MgO in the whole rock, the hypersthene as well as the biotite carries almost exactly 0.50 FeO:0.50 MgO and the ilmenite 0.96 FeO:0.04 MgO.

The proportions of FeO and MgO in the ilmenite, from observation of the three rocks just mentioned, must be a function of the FeO:MgO proportion in the entire rock or in the original magma, but in such a manner that the ilmenite throughout shows a very extensive relative enrichment of FeO (as $FeTiO_3$) and consequently vice versa an extensive, relative decrease of MgO (as $MgTiO_3$).

[*To be continued*]

CYCLES OF EROSION IN THE PIEDMONT PROVINCE OF PENNSYLVANIA[1]

F. BASCOM
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Since 1912, when Professor Barrell brought to the attention of the Geological Society of America some conclusions opposed to the earlier interpretation of the erosion history of certain portions of the Appalachian highlands, the writer has had in mind the possible application of similar conclusions to the erosion history of the Piedmont province of Pennsylvania. The results of this intention are presented in this paper. The conclusions reached are not precisely in accord with those enforced with so much originality by Professor Barrell, nor do they involve much that is new in the interpretation of the erosion history of eastern Pennsylvania, but they are presented as a record of the present stage of the study of the peneplains of the Piedmont province of Pennsylvania.

It is the purpose of the paper to call attention to the fact that the erosion history of eastern Pennsylvania as indicated by altitudes and by the record of sedimentation must have been complex, that it was made up not of one or two or three cycles of prolonged erosion, but of many interrupted cycles, and that vestiges of nine of these cycles testify to their reality. Other cycles may have existed and probably did exist, but too briefly for permanent record. Six of these nine cycles are thought to belong to post-Cretaceous time and three to Cretaceous time.

The question of the subaerial or marine origin of these peneplains is debated, but decisive criteria are lacking for a final pronouncement.

[1] Published by permission of the Director of the United States Geological Survey. The writer takes pleasure in acknowledging her indebtedness to M. R. Campbell and G. W. Stose of the United States Geological Survey for helpful comments and queries made on the subject-matter of this paper, and to Professor W. M. Davis for valuable specific suggestions.

In any investigation of the cycles of erosion, complete or incomplete, that collectively constitute the erosion history of the Piedmont province of the Appalachian highlands, the stratigraphic record preserved on the margin of the province must furnish the data by which the succession and age of such erosion cycles stand or fall.

That aerial and marine erosion has taken place ever since continental plateaus and oceanic basins came into existence is unquestioned: only such interaction of air, water, and land masses is conceivable. The character and rapidity of erosion will be controlled by altitude, rock, and climate, but the duration of an erosion period will be dependent upon the stability of the strand line: a long period of quiescence will permit prolonged aerial and marine erosion with reference to a given base-level, and a period of uplift will interrupt and renew erosion with reference to a new base-level. The evidence of such movements of the strand line inaugurating erosion is furnished by the stratigraphic register.

In the Piedmont province of Pennsylvania with the beginning of Cretaceous sedimentation the stratigraphic record seems to indicate a succession of such erosional conditions maintained by an alternation of periods of continental quiescence with periods of movement. That these periods of stability have been of different durations is an obvious deduction from the sedimentary record.

The stratigraphic record on the Atlantic plain which is the submerged margin of the Piedmont province is as follows:

Recent deposits
  Unconformity
Pleistocene deposits
    Talbot (Cape May): clay, sand, and gravel..................30 feet
  Unconformity
    Wicomico (Pensauken): clay, sand, and gravel................25 feet
  Unconformity
    Sunderland (Bridgeton): clay, sand, and gravel...............25 feet
  Unconformity
    Late Brandywine: sand and gravel.........................1± foot
Pleistocene (or Late Tertiary ?) deposits
  Unconformity ?
    Early Brandywine: sand and gravel........................50 feet

Miocene deposits	
Unconformity	
St. Mary's: sand and clay	280 feet
Choptank: sand, clay, and marl	175 feet
Unconformity	
Calvert: sand and clay	310 feet
Unconformity	
Eocene deposits	
Nanjemoy: sand	125 feet
Aquia: greensand	100 feet
Unconformity	
Upper Cretaceous (Cretaceous) deposits	
Manasquan: clay and sand	50 feet
Rancocas: greensand	80 feet
Monmouth: sand	100 feet
Unconformity	
Matawan: micaceous sandy clay	70 feet
Unconformity	
Magothy: sand and clay	100 feet
Unconformity	
Raritan: clay and sand	350 feet
Unconformity	
Lower Cretaceous (Comanchean) deposits	
Arundel: clay and sand	125 feet
Unconformity	
Patapsco: clay and sand	200 feet
Unconformity	
Patuxent: sand and arkose	350 feet
Unconformity	
Crystalline formations	

With the Cretaceous, Tertiary, and Pleistocene registration of continental movements before us, it is no longer possible to believe that the erosion history of this region is told in two cycles of erosion, producing two peneplains: the Kittatinny,[1] or Schooley, of Cretaceous age, and the Shenandoah, or Somerville,[2] of Tertiary age. That there is topographic evidence of more than two erosion

[1] Bailey Willis, "The Northern Appalachians," *Nat. Geog. Mon.*, 1895, pp. 169–202. C. W. Hayes, "The Southern Appalachians," *Nat. Geog. Mon.*, 1895, pp. 305–36. C. W. Hayes and M. R. Campbell, "Geomorphology of the Southern Appalachians," *Nat. Geog. Mon.*, 1894, pp. 63–126.

[2] W. M. Davis and J. W. Wood, Jr., "The Geographic Development of Northern New Jersey," *Proc. Bost. Soc. of Nat. Hist.*, Vol. XXIV (1889), pp. 365–423.

periods in the Appalachian highlands has been recognized by Keith,[1] Campbell,[2] and others.

In the sedimentary sequence of the Atlantic plain there are ten significant unconformities—that is, ten intervals of erosion alternating with intervals of deposition. Not all of the deposits are known to be marine, so that ten submergences cannot be postulated. There are six less significant unconformities.

The time represented by the deposits and unconformities has been estimated at 56,500,000 years.[3] There could not conceivably be conditions more favorable for a succession of erosion cycles falling so far short of completion as to leave permanent traces of the sequence, nor a stratigraphic record more compelling for the acceptance of such traces as evidences of erosion cycles. It is not probable that traces are preserved of every incomplete erosion cycle.

The topography of erosion cycles early interrupted would be obliterated by subsequent erosion cycles of longer duration. Small beginnings, if they existed, might be quite similar to the three most recent terraces, which are being modified and will in time be completely obliterated by subsequent erosion.

Such incomplete, obliterated cycles may be registered only in the lesser unconformities of the stratigraphic record, which is easily more complete than the topographic record. Did the geologist base his expectations on stratigraphy alone, he would look for a series of more or less discontinuous and more or less warped benches or terraces facing the sea, or, in the case of the lower terraces, following inland the river valleys, and not perfectly stairlike because each terrace will have its peculiar angle of slope. The terraces are the topographic record of the succession of interrupted erosion cycles of which the unconformities in the stratigraphic sequence are the geologic record.

[1] Arthur Keith, "Some Stages of Appalachian Erosion," *Bulletin Geol. Soc. America*, Vol. VII (1896), pp. 519–24. "Geology of the Catoctin Belt," *Fourteenth Annual Report, U.S. Geol. Sur.*, Part II (1892–93), pp. 285–395.

[2] M. R. Campbell, "Geographic Development of Northern Pennsylvania and Southern New York," *Bulletin Geol. Soc. America*, Vol. XIV (1903), pp. 277–96.

[3] Joseph Barrell, "Rhythms and the Measurement of Geologic Time," *Bulletin Geol. Soc. America*, Vol. XXVIII (1917), pp. 745–904.

Nature is less obvious and more complex in her methods and evidences than such expectations would imply, but a detailed and careful study of approximately level tracts and benches seems to justify the following series of peneplains and terraces related to the major unconformities of the stratigraphic record.

Name	Altitude West East	Sediments	Age	Preserved
Peneplains				
Kittatinny.....	1800–1600–1100	Patuxent.	Jurassic and Lower Cretaceous	Quartzite
Schooley......	1300–1000–900	Patapsco-Arundel.	Lower Cretaceous	Granite
Honeybrook...	860–800–700	Raritan-Manasquan.	Upper Cretaceous	Granite
Harrisburg	800–500	Aquia-St. Mary's.	Tertiary	Shale
Early Brandywine....	500–400–390	Early Brandywine.	Pliocene (Pleistocene)	Shale, etc.
Terraces				
Late Brandywine ...	400–300–200	Late Brandywine.	Pleistocene	Mica gneisses
Sunderland....	300–180–100	Sunderland.	Pleistocene	Mica gneisses
Wicomico.....	90–45	Wicomico.	Pleistocene	Mica gneisses
Talbot........	45–40–0	Talbot.	Pleistocene	Mica gneisses

The oldest peneplain, the highest inland from the sea and the lowest near the sea where it is preserved under sedimentary rocks, is the *Kittatinny*. This peneplain surface is so strikingly upheld,

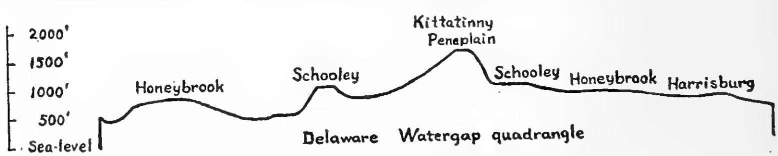

FIG. 1.—Section across Godfrey Ridge and Kittatinny Range, at the Delaware Water Gap, Delaware Water Gap quadrangle, Pennsylvania–New Jersey.

although not preserved unmodified, by the indurated sandstone of Kittatinny Mountain (1,600 feet), its type locality (see Fig. 1), and in other resistant ridges of the Appalachian highlands that it was early recognized and an effort was made to fit it to the lower summits of Schooley Mountain and still lower surfaces in the

Piedmont province. In order to secure continuity between adjacent and discordant levels, it was necessary to assume abrupt and steep warping of the peneplain in some localities, and when accordance was secured it left unexplained the remarkable preservation throughout the Piedmont province of so ancient an erosion surface, and failed to explain why no records were preserved of the later continental movements and erosion cycles which are recorded in the sedimentary succession.

FIG. 2.—Kittatinny, Schooley, and Honeybrook peneplains in the Reading quadrangle. The summit of the ridge on the left at 1,140 feet represents the Kittatinny; the ridge in the middle at 1,000 feet, the Schooley; the ridge on the right at 700 feet, the Honeybrook peneplain. West Reading in middle distance, looking east.

The Kittatinny peneplain has been traced northeastward from the type locality to the base of the Catskill Mountains in New York, and westward and southward into Maryland and West Virginia. In the Blue Ridge province, surfaces which are probably remnants of the Kittatinny have altitudes of 1,800 feet in southern Pennsylvania (South Mountain), 1,300 on Blue Mountain to the northeast, and 1,200 feet in the quartzite ridge east of Reading (Penn Mountain, the dominating highland of the area, designated the Reading Prong of the New England upland). (See Figs. 2 and 3.)

Whether the Kittatinny peneplain is anywhere preserved in the Piedmont province of Pennsylvania is questionable. Reduced remnants of it may appear on Welsh Mountain (Honeybrook

quadrangle), a quartzite ridge, but, as is to be expected in a region so near the sea, erosion in subsequent cycles has probably modified the Kittatinny surface, notwithstanding the resistant character of the rock.

Near the "fall-line" where the lowest and oldest formations (Patuxent formation) of the coastal plain lie directly upon a peneplained surface of crystalline rocks, the floor which bears them may be that part of the Kittatinny peneplain which was submerged, was buried beneath sediments, and was thus preserved without modification while far inland the peneplain was still developing.

FIG. 3.—Schooley peneplain in the Reading quadrangle. The summit of Irish Mountain in the distance at 1,000 feet represents the Schooley peneplain, as seen from a point on the Early Brandywine peneplain, two miles east of Shoemakersville, looking south 45° east.

The Patuxent formation once overlapped the margin of the Piedmont province to a distance inland considerably greater than is now covered by it; but wherever it has been removed by erosion, the surface of the peneplain has been attacked so that the old surface cannot be found except perhaps in the immediate vicinity of the remnants of the formation. This surface lies at about 180 feet above sea-level, rising in Maryland to 280 feet. This first peneplain, carved on a dissected highland or possibly on uplifted peneplains, obliterated in this region all pre-existing erosion surfaces except those that were protected by a cover: an example of such a surface is to be found on Paleozoic rocks, where they are covered by Triassic formations (see Fig. 4). The later peneplains, carved on uplifted peneplains or terraces, never completely obliterated pre-existing erosion surfaces.

That the Kittatinny erosion cycle exceeded in duration any of the subsequent cycles must have been the case not alone because no subsequent cycle has been coextensive with it, but also because no subsequent cycle has succeeded in wearing down the most resistant rocks leveled by Kittatinny erosion and located well within the area of subsequent peneplanation.

The next oldest peneplain, the Schooley, with its type locality the granite summits of Schooley Mountain (1,300 feet), New Jersey, has been traced northward to the Mohawk Valley, westward to

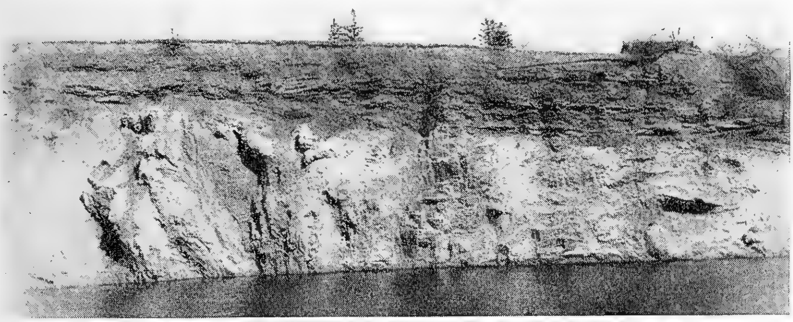

FIG. 4.—Section of erosion surface on the Paleozoic, protected by a cover of Triassic rocks. Port Kennedy, Montgomery County, Pennsylvania.

Syracuse,[1] and southward to the Potomac.[2] In western New York it appears to coalesce with the Kittatinny, suggesting that in that region there was no uplift separating the two erosion periods.

At the Delaware Water Gap[3] the ridge between Godfrey Ridge and Kittatinny Range with summit areas at 1,000 feet may represent the Schooley peneplain, and Wind Gap between Blue Mountain and Kittatinny Mountain in Northampton County at the same elevation may have been a water gap during the early part of the Schooley erosion cycle. In the Blue Ridge province remnants are preserved in summit areas of 1,200 and 1,000 feet

[1] Memorandum by M. R. Campbell.

[2] Mem. by the writer.

[3] G. W. Stose, "Text of Delaware Water Gap Sheet," *U.S. Geol. Survey.*

altitudes. East and northeast of Reading there are many flat-topped granite and quartzite hills rising to a height of 1,000 feet, the Schooley level in that locality (see Figs. 5–10). On one of these, the Schooley remnant is separated from the adjacent Kittatinny

FIG. 5.—Schooley peneplain above the Honeybrook peneplain in the Boyertown quadrangle. The summit of Long Hill in the distance at 1,040 feet represents the Schooley peneplain, as seen from a point one-half mile southeast of Shanesville, looking south. The summit of the ridge in the foreground is a remnant of the Honeybrook peneplain.

FIG. 6.—Schooley peneplain in the Reading quadrangle. The higher parts of the past-maturely dissected upland one and one-half miles southeast of Fleetwood represent the Schooley peneplain at an altitude of 940–1,000 feet. Hill road, looking south on hills south of Princeton.

remnant by a steep slope (Fig. 9). In the central Piedmont province the Schooley peneplain descends to an altitude of 800 feet (Coatesville quadrangle). If the Schooley peneplain reappears near the "fall-line," it is found on the border of, and passing

beneath, the Patapsco formation, which rests upon eroded Patuxent, at an altitude of 100 feet, rising to 130 feet in Maryland. The next movement of uplift not only raised the Schooley peneplain

FIG. 7.—Schooley, Honeybrook, and Late Brandywine peneplains in the Boyertown quadrangle. High hills in the background are remnants of the Schooley peneplain, altitude 1,000 feet; hill in center in middle distance is at the level (660 feet) of the Honeybrook peneplain; and the foreground is on the Late Brandywine peneplain, altitude 420 feet. Looking west from Palm Station.

FIG. 8.—Schooley and Honeybrook peneplains in the Boyertown quadrangles. Hills in the background are remnants of the Schooley peneplain, altitude 1,000 feet; foreground on the Honeybrook peneplain, altitude 800 feet. Devil's Hump, looking south 30° west.

and remnants of the Kittatinny peneplain to a considerable height, but warped them.[1]

The next younger peneplain, the Honeybrook, appears in Godfrey Ridge, northwest of Kittatinny Mountain, at the Delaware

[1] Bailey Willis, *op. cit.*, pp. 189–90. C. W. Hayes, *op. cit.*, p. 330.

Water Gap, and on the hill summits southeast of Kittatinny Mountain at an altitude of 800 feet. This altitude is a very persistent one in the Appalachian Valley from this region to Susquehanna River. The Hamburg and Slatington quadrangles show the Honeybrook peneplain dominating the interstream areas. It retains an altitude of 800 feet in the Blue Ridge province and is well shown east of Reading, where Neversink Mountain, Guldin Hill, and the southeastern spur of Penn Mountain preserve its surface (see Fig. 9). The Honeybrook and the Schooley are here

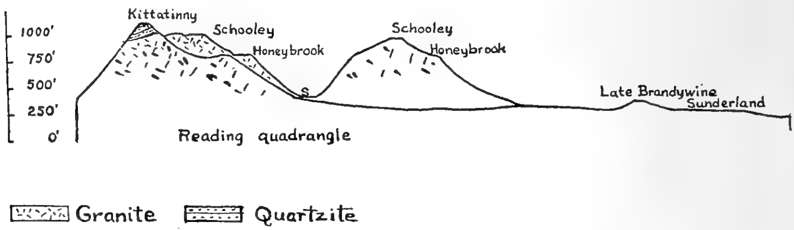

FIG. 9.—Section across Penn Mountain and Guldin Hill, showing remnants of the Kittatinny, Schooley, and Honeybrook peneplains, and of the Late Brandywine and Sunderland erosion surfaces. Reading Prong of the New England upland, Reading quadrangle, Pennsylvania.

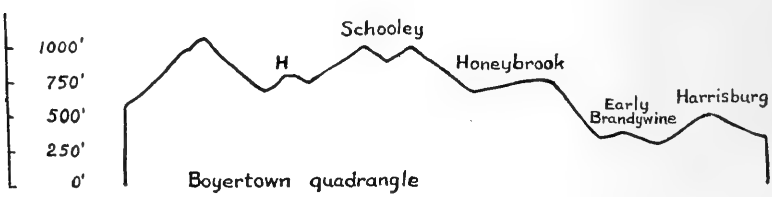

FIG. 10.—Section across Long Hill, Devil's Hump, and Gabel Hill, showing the Schooley, Honeybrook, and Harrisburg peneplains. Boyertown quadrangle, Pennsylvania.

found adjacent, are both cut in granite, and are separated by a steep slope (see also Figs. 11 and 12).

In the Piedmont province the Honeybrook descends to 700 feet. On the divide between Susquehanna and Schuylkill rivers the North and South Chester Valley Hills preserve this peneplain. The most extended remnant of it is found on the granite about Honeybrook, 16 miles south of Reading, and from this type locality

the peneplain is here named the Honeybrook.[1] It is not claimed that this plain has been traced throughout the Appalachian highlands division, but in the Piedmont province of Pennsylvania it

FIG. 11.—Honeybrook peneplain below the Schooley, Boyertown quadrangle. The upland in the distance represents the Schooley peneplain, altitude 1,000 feet, and that in the middle distance, altitude 800 feet, the Honeybrook peneplain, as seen from a point one-fourth mile southwest of Shanesville, looking north 15° west.

FIG. 12.—Water gap in a ridge whose summit is a remnant of the Honeybrook peneplain, Boyertown quadrangle. Upland in background, altitude 1,000 feet, represents the Schooley peneplain. Summit of ridge 800 feet and stream in the water gap, 440 feet. View from a point one-fourth mile southwest of Shanesville, looking north 45° west.

seems to represent a distinct erosion level between the Schooley and Harrisburg. The Honeybrook peneplain has been completely

[1] The Schooley peneplain was traced from Pennsylvania to the Potomac Valley in Maryland to surfaces (Green Ridges) which have been ascribed to the Weverton peneplain (*Maryland Geol. Survey*, Vol. VI [1906], pp. 87–88). Elsewhere in the central Piedmont of Pennsylvania the Weverton as defined corresponds to a lower peneplain than the Schooley. A new name has therefore been introduced for a redefined Weverton.

removed at the "fall-line." It passes under the Raritan (Upper Cretaceous) formation near the junction of the Coastal Plain and Piedmont.

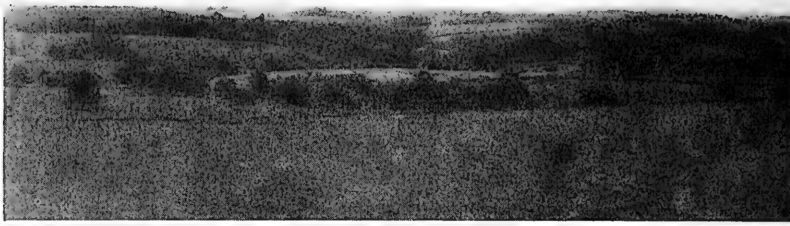

FIG. 13.—Harrisburg peneplain in the Coatesville quadrangle. The dissected peneplain at an altitude of 600 feet, as seen one-third of a mile northwest of Humphreyville, looking southwest.

FIG. 14.—Harrisburg peneplain on Schuylkill River, in the Reading and Honeybrook quadrangles. The summit of the hill in the middle distance in which the river is cutting a steep bluff represents the Harrisburg peneplain, altitude 600–660 feet. Gibraltar Hill, 900 feet, a monadnock, on the left, and the Sunderland plain in the foreground, as seen from Lookout Point, Neversink Mountain, looking south.

The Harrisburg peneplain[1] has been restricted, with the approval of its sponsor, at its type locality northeast of Harrisburg to upland

[1] M. R. Campbell, *Bulletin Geol. Soc. America*, Vol. XIV (1903), pp. 277–96.

surfaces on the Ordovician (Martinsburg) shale which reach 600 feet, and to corresponding altitudes on Delaware and Potomac rivers (see Figs. 13 and 14). At this altitude it is widespread in

FIG. 15.—Honeybrook and Harrisburg peneplains in the Boyertown quadrangle. The Honeybrook peneplain corresponds with the surface of the upland on the left at an altitude of 800 feet and the Harrisburg with the upland in the distance on the right at an altitude of 600 feet, as seen from the northeast end of Long Hill, looking north 55° east.

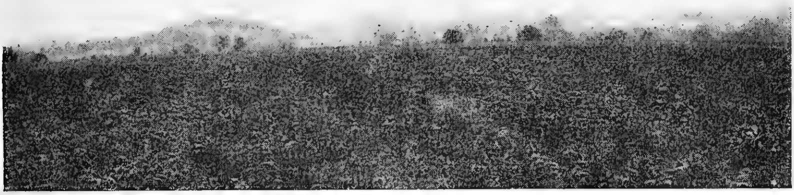

FIG. 16.—Remnants of Early Brandywine and Harrisburg peneplains in the Harrisburg quadrangles. The surface of the upland represents the Early Brandywine peneplain upon which the hills, rising to an altitude of 740 feet and perhaps to the level of the Harrisburg peneplain, stand as monadnocks. The view is from a point on the Sunderland level one-half mile northwest of Maiden Creek, looking southwest.

the central Piedmont province and descends on the border of the upland to 500 feet (North and South Chester Valley Hills in the Schuylkill Valley). The Harrisburg does not appear in the "fall-line" zone, but probably descends below sea-level beneath the

Aquia Greensands (Tertiary), which lie far out on the Coastal Plain.

The Early Brandywine, the youngest and most widely preserved of the five peneplains, is found on Ordovician shale at the 500-foot level, northeast of Harrisburg. It contains at this altitude in the

FIG. 17.—Early Brandywine peneplain in the Boyertown quadrangle. The peneplain corresponding with the surface of the upland in the distance at an altitude of 560 feet is seen from a point one-half mile west of Eschbach, looking south 45° east.

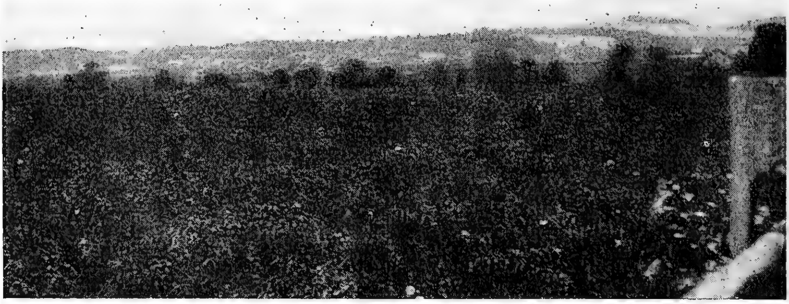

FIG. 18.—Early Brandywine peneplain in the Reading quadrangle. The peneplain at an altitude of 500 feet is represented by the surface of the upland in the distance, as seen from a point on the Sunderland level one-half mile northwest of Maiden Creek, looking south 65° west toward Leesport.

valley of the Delaware at the Water Gap and in the Schuylkill and Potomac valleys (Antietam quadrangle). It ranges from 400 to 450 feet in the Piedmont upland of Pennsylvania where, as is to be expected, because it is the most recently formed peneplain, it is the most pronounced upland level (see Figs. 15–18).

West Chester is located upon this peneplain surface, which is the dominant altitude throughout the West Chester quadrangle (see Figs. 19 and 20). At 400 feet it carries Early Brandywine gravel and sand, 10 miles west of the "fall-line" zone. It has been named the Early Brandywine from the formation which is found at this altitude and on the seaward continuation of the slope. This peneplain is correlated with the so-called Lafayette[1] terrace recognized in Maryland[2] but a more widespread extension is claimed for the Early Brandywine peneplain.

The Early Brandywine peneplain is everywhere submaturely dissected. The summits of the interstream areas preserve the peneplain, gentle slopes from these summit remnants lead to the gorges (Pleistocene) of the main streams and of the larger tributaries or form the U-shaped valleys of headwater streams. These slopes have an elevation inland from 300 to 400 feet and in the "fall-line" zone from 200 to 300 feet (see Figs. 21–25).

Following the deposition of the Early Brandywine formation and before the deposition of the Sunderland formation, the whole continental shelf was brought above the sea and master-streams of the Atlantic plain were extended to the edge of the continental shelf. To this period, which may have been well within Pleistocene time, is attributed the formation of the Late Brandywine benches and slopes. Few formations can be correlated with it, as the

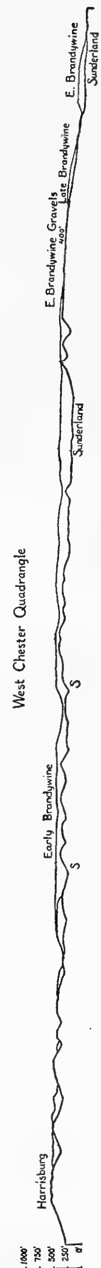

FIG. 19.—Section in West Chester quadrangle

[1] Owing to the change of the name Lafayette to Brandywine, the more recent name has been given to the peneplain. The name Brandywine is taken from a village of that name in Prince George County, Md., where the formation is reported to be characteristically developed. The position and level of the gravel of this type locality at 233 feet seem to indicate that it is the low-level Brandywine or Late Brandywine gravel as it is provisionally named in this paper.

[2] *Maryland Geol. Survey*, Vol. VI (1906), pp. 59–60.

marine sedimentation of the period took place, mainly at least, beyond the continental shelf. Gravel, which has been included in the "Brandywine" (Early and Late Brandywine), but which lies at all places at a lower level than the Early Brandywine gravel, is thought to be a terrestrial deposit of Pleistocene streams. Such gravel is found on the Chester quadrangle at an altitude of 300 feet and on Elk Neck, Elkton quadrangle, between 200 and 300 feet.

The records of this period of erosion are the dissection of the Early Brandywine peneplain, producing the stream terraces and

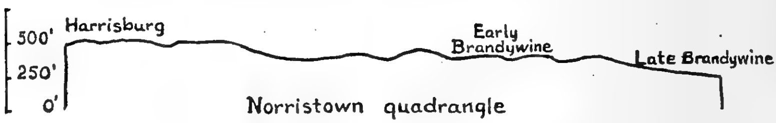

FIG. 20.—Section in Norristown quadrangle

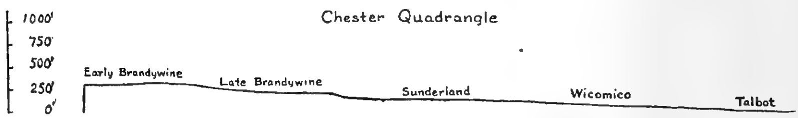

FIG. 21.—Section in Chester quadrangle

the slopes which separate the Early Brandywine peneplain and the Sunderland terrace, and the submerged valleys on the continental shelf. Late Brandywine slopes are well defined on the Chester quadrangle, and furnish a commanding site for the buildings of Swarthmore College.

The Sunderland, Wicomico, and Talbot terraces have been recognized and defined in Maryland.[1] In Pennsylvania a scarp separates the Late Brandywine and the Sunderland. This scarp, which the central building of Swarthmore College fronts, represents either the old estuarine shore cliff or the escarpment of the wide meander belt of Delaware River.

Erosion truncated the Late Brandywine slopes and dissected them and the Early Brandywine peneplain along drainage ways. What has been called the Somerville peneplain seems to the writer

[1] *Maryland Geol. Survey*, Vol. VI (1906), pp. 61–67.

to be such an inland extension of erosion during the Sunderland cycle.

In general the Sunderland extends from the 100 to the 180 contour lines: the Wicomico from the 80 to the 90 contour lines, and the Talbot, where it does not coalesce with the Wicomico, from the 40-foot contour to sea-level. These three terraces are conspicuously developed in eastern Pennsylvania parallel to Delaware River. In Maryland the Wicomico and Talbot terraces are in some places obliterated and the Sunderland reaches the edge of the beach with a cliff 100 feet high, but this is not the case in Pennsylvania where the terraces are not seacoast features.

The Wicomico terrace wraps about the Sunderland as the Sunderland does about the Late Brandywine, with usually a well-marked break between the two, except in the gorges of the tributary streams. The Talbot terrace borders the Wicomico, which it penetrates along drainage ways, and in some places parallel to Delaware River coalesces with the Wicomico.

It has not proved practicable to show by graphic means the distribution of the remnants of the peneplains and terraces in the Piedmont province of Pennsylvania. It may be stated that in general the oldest peneplain is farthest inland and the youngest nearest the shore, with those of intermediate age ranging between. If this region had been one of uniform resistance to weathering, there would have been a perfect operation of this law of areal

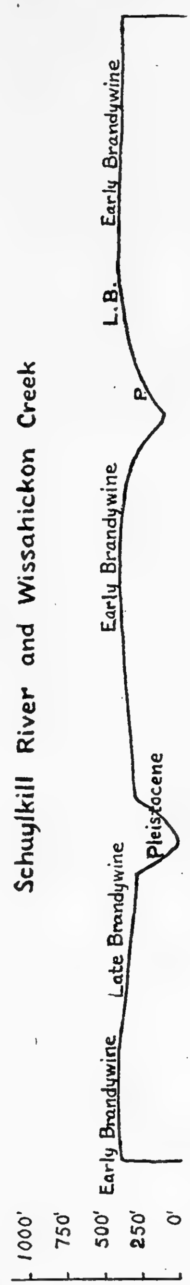

FIG. 22.—Section crossing Schuylkill River and Wissahickon Creek

distribution: the areal succession of peneplains from interior to coast would exactly accord with the chronological succession. The region is, however, one of varied structural and lithologic resistance to weathering and the peneplains are not therefore so simply spaced; younger peneplains on relatively weak rocks are found inland at higher altitudes than the marginal remnants of older peneplains. This fact would be still more apparent if the extreme margins of the older peneplains, now buried beneath sedimentary formations, were shown.

The question of the origin of these peneplains, that is, of the nature of the dominant erosive agent, is open to debate. The

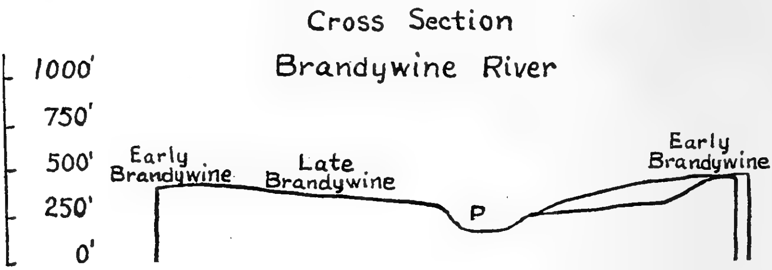

FIG. 23.—Section crossing Brandywine River

three youngest terraces, in Maryland presumably of marine origin, are in this region of fluvial-estuarine or of fluvial origin; that is, they were developed on the borders of the Delaware estuary or on a shrinking meander belt of Delaware River.

That the Late Brandywine is of subaerial origin is concluded from the evidence of the valleys, now submerged, which extend across the continental shelf and which it is believed were excavated in Late Brandywine time.

That the five peneplains are in part of marine and in part of subaerial origin seems a warranted conclusion. Each peneplain was partly submerged and carries marine sediments, but there does not seem to be sufficient proof that any one peneplain was completely submerged. They parallel the coast line as would be the case were they of marine origin, but this may also be true of subaerial peneplains, and the great inland extension of the Kittatinny

and Schooley peneplains with an indefinite thin margin is indicative of subaerial erosion. The contact of the Honeybrook and Schooley peneplains, on the Reading quadrangle, on the other hand, suggests

FIG. 24.—Late Brandywine peneplain in the Reading quadrangle. The peneplain at an altitude of 400 feet, as seen from the hillside south of Oley Furnace, looking south toward Friedensburg.

FIG. 25.—Honeybrook and Late Brandywine peneplains in the Reading and Boyertown quadrangles. The surface of the upland in the distance at an altitude of 800 feet represents the Honeybrook peneplain, and the level land in the middle distance at altitudes ranging from 400 to 440 feet, the Late Brandywine peneplain, as seen from a point one-fourth miles southwest of Oley Furnace, looking south 45° east toward Shenkel Hill.

a sea cliff. In the case of the Harrisburg and Early Brandywine peneplains definite proof of subaerial or marine origin has not been found.

THE HORIZONTAL MOVEMENT OF GEANTICLINES AND THE FRACTURES NEAR THEIR SURFACE

H. A. BROUWER
Delft, Holland

Most islands of the arcs which lie to the east and the southeast of the Asiatic continent show proof of an uplift of the land relatively to the sea-level, which is amply demonstrated in tropical regions by the presence of upheaved fringing reefs. In the East Indian Archipelago there exists a striking difference between the western and the eastern parts as regards the rising islands and the submarine topography. If the sea-level were to be lowered 200 m., Sumatra, Java, and Borneo would form one mass of land with the peninsula of Cambodia and Siam, just as Australia would form a single mass with the Aru Islands through the vast tract now occupied by the shallow Arafura Sea and the Bay of Carpentaria to New Guinea and the islands Misool, Waigeu, Batanta, and Salawati to the west of New Guinea.

Between these two near-land-masses lies an area in which deep sea basins alternate with upheaved islands. From a geological point of view Verbeek[1] first drew attention to this remarkable fact, of which a more satisfactory discussion has been made possible because of the new deep-sea chart of the Siboga Expedition.[2] In Verbeek's opinion the elevation of the islands surrounding the Banda Sea is the result of folding at greater depth. The active forces first began compressing near the surface, and as the geosynclines were formed they became active at greater depths. Later Molengraaff[3] expressed similar ideas, and for the southeastern por-

[1] R. D. M. Verbeek, "Rapport sur les Moluques," édition française du *Jaarb. v. h. Mijnwezen in Ned. O. Indië,* Vol. XXXVII (1908), pp. 833, 834.

[2] G. A. F. Tydeman, "Hydrographic Results of the Siboga Expedition," Chart 1, in M. Weber, *Siboga-Expeditie,* Part III, Leyden, 1903.

[3] G. A. F. Molengraaff, "Folded Mountain Chains, Overthrust Sheets and Block-Faulted Mountains in the East Indian Archipelago," *Compte rendu du XII^e congrès géologique international,* Toronto, 1913, p. 699.

tion of the Malay Archipelago he distinguished two types of mountain-building: (1) the overthrust type of Miocene age, culminating in overthrusts of great magnitude, which was the expression of a very powerful, but not deep-seated compression, and (2) the block-faulted type of Plio-Pleistocene age, consisting of ranges of elevated islands alternating with deep sea basins, these being the expression of a deeper-seated, but perhaps less energetic compression.

It is only the vertical movements of the rows of islands that have been considered by these authors. In some recent publications[1] I have pointed out that:

1. The youngest crustal movements in this region are a younger phase in the same process as the older and an exact continuation of the mid-Tertiary crustal movements. Of the mid-Tertiary phase we know only the folds and overthrusts which represent action at greater depth; of the youngest phase only the fractured and faulted crust which represents action near the surface; but the two phenomena are mutually complementary and the rows of uplifted islands indicate the spots where the folding process continues at the greater depths with the same tendency to form overthrusts.

2. From the outline of the rows of islands we may conclude that they have a large movement in a *horizontal* as well as in a vertical direction.

The horizontal movements of the curving rows of islands are expressed by several of their characters.

1. The striking fact that the Tenimber Islands and the Kei Islands have an outlying position in the row and both are situated opposite a depression in the Sahul bank which constitutes the Australian continental shelf. Opposite these depressions the geanticline met with less resistance.

[1] H. A. Brouwer, "On the Crustal Movements in the Region of the Curving Rows of Islands in the Eastern Part of the East Indian Archipelago," *Proceed. Kon. Akad. v. Wetensch. Amsterdam*, Vol. XXII, pp. 772–82; "On Reef Caps," *ibid.*, Vol. XXI, pp. 816–26; "Fractures and Faults near the Surface of Moving Geanticlines," *ibid.*, Vol. XXIII, pp. 570–76; "Über Gebirgsbildung und Vulkanismus in den Molukken," *Geol. Rundschau*, 1917, p. 197; "Über die horizontale Bewegung der Inselreihen in den Molukken," *Nachr. d. Gesellsch. der Wiss. zu Göttingen*, 1920.

2. The coincidence of asymmetrical reef caps with marked outward bends of the row of islands, instances of which are found in the island Rotti to the southwest of Timor and in the island Jamdena of the Tenimber group.

3. The faults and fractures near the surface demonstrate differences in rate of horizontal movement between adjacent parts of the moving geanticlines.

In the following pages the above-described faults and fractures will be dealt with in connection with the vertical and horizontal movements of the geanticlines near the surface of which they occur. Because the geanticlines have risen from the sea and were in consequence exposed to eroding influences during a much shorter time than those of the continental mountain ranges, the outer form is not in the main controlled by erosion, but by the crustal movements themselves, and the latest phase of mountain-building manifests itself clearly in the shape of the geanticlines near the surface.

CRUSTAL MOVEMENTS AND MORPHOLOGICAL STRUCTURE

When crustal movements take place they generally cause the strata to break near the surface and to fold at greater depths. An extension of the geanticlinal axis is here obtained through gaping fractures, or by movements parallel to fault planes which must be inclined to the geanticlinal axis. Shortening of the geanticline is possible by faulting along fault planes which are not perpendicular to the geanticlinal axis. Similar relations prevail for a lengthening or a shortening of a section of the geanticlinal surface with a plane perpendicular to the geanticlinal axis.

In addition to the control by the direction and the rate of the movement, the position of the fault planes is determined by a great many other factors, e.g., by stratification and by the composition and distribution of the rocks near the surface. Leaving out of consideration those local areas within which the anticlinal axis shows an important pitch, the morphological aspect of the surface will be controlled chiefly by the more or less horizontal transverse faults, the gaping transverse fractures, the more or less longitudinal faults, and the gaping longitudinal fractures.

We are here considering those regions only of the geanticlinal surface where the faults, through their more or less equal position

and their more or less equal direction of movement, bring about considerable alterations in the broad outlines of the morphological structure. Zones of constant lithological characters will generally be separated near the surface by planes which are parallel to the geanticlinal axis. If these planes are more or less vertical, this will chiefly influence the distribution of the vertical longitudinal fractures and the longitudinal faults. If these planes are principally more or less horizontal, this will chiefly influence the distribution of the faults along horizontal planes, but they will be of little importance for the major morphological structure and will here be left out of consideration. Whether these planes are nearly vertical or nearly horizontal, the lithological character is of little importance for the distribution of the transverse faults and fractures which strongly influence the morphology at the surface of the geanticline. Thus we find that the outline at the surface is mainly controlled by the direction and the rate of the crustal movements in so far as the transverse fractures are concerned.

OLDER FOLDS CUT OFF BY THE PRESENT COAST LINE

The surface and the deeper parts of moving geanticlines will generally not move in the same direction and at the same rate, because:

1. The intensity and likewise the direction of the forces which cause the movement near the surface will generally be different from those which obtain at greater depth.

2. The transmission of directed forces will decrease from the surface to the zones of higher plasticity at greater depth.

If the forces which cause the movement are deep-seated, and the crust near the surface does not respond to the direct influence of the compressional or tensional stress, the displacements near the surface will be the result of the movement at greater depth. In forming a judgment on the genesis of fractures and folds this should be borne in mind.

A result of the difference between the movements at greater depth and those near the surface is that, if at greater depth the movement has a horizontal component, those points which were originally on the same vertical line will in a later stage of evolution of the geanticline form an irregular curve, the form of which will

depend upon the direction and the rates of movement at different depths. If a geanticline is elevated above the sea, the deeper-seated parts will gradually be uncovered by erosion and the surface of the geanticline will in time consist of rocks which were in the zone of flow during an earlier stage of the mountain-building process. As they are approaching the earth's surface, the rate and the direction of the motion may differ more and more from those at greater depths on the same vertical line.

That older folds terminate abruptly against the present coast lines is a phenomenon which is well known from Japan and from several islands of the East Indian Archipelago (Fig. 1). Particularly on Ceram this fact is very strikingly exemplified. In the

FIG. 1.—Older folds terminating abruptly against the present coast lines of the island of Ceram. (East Indian Archipelago.) Scale 1:3,000,000. ------ Approximate Tertiary strike.

greater part of the island the strike of the Tertiary mountain range is NW.–S.E., whereas the present coast line has for the middle part an east–west direction, so that the ridges of the high mountains terminate abruptly near Taluti Bay on the south coast and near Savai Bay on the north coast.

Similar facts have been explained by von Richthofen[1] as a result of tensional stress on a large scale, and he believed that the mountain arcs of eastern Asia, although bearing a great resemblance to the Alps and the Himalayas, have been formed by tensional, and not by compressional stress. Various authors have pointed out that this conception is not exact, and particularly because the fractures resulting from tensional stress are generally straight, whereas the ranges which lie to the eastward of the

[1] F. von Richthofen, "Geomorphologische Studien aus Ost-Asien, IV," *Sitzungsber. der Berlin. Akad. der Wiss.*, XL (1913).

Asiatic continent are arcs which present their convex sides to the oceanic areas. The tension hypothesis of Von Richthofen has been applied by some authors to the East Indian Archipelago, but the numerous fractures which without doubt exist near the surface can be explained in a simpler manner by the action of compressional stress.

It is not necessary to distinguish two periods of folding with different directions of the compressive forces, if we have regard for the fact that the older folds are cut off by the present coast lines. If the strike of the older folds is independent of the outlines of the present rows of islands, this may be in part a result of a change in the direction of the compressive forces; but it can be entirely a result of the fact that the folds which now appear at the earth's surface have been formed in a much earlier stage of evolution of the geanticline, and that during their elevation the horizontal component of the rate of movement was different for neighboring parts of the geanticline, while the transmission of the directed forces has increased and the intensity and the direction of the forces has changed, whereas at greater depths the plastic deformation has continued.

GROUPS OF SMALL ISLANDS WITH HIGH REEFS

In many rows of islands the breadth of each island is in direct proportion to the amount of elevation. In the Timor-Ceram range the long and broad island of Timor shows elevated reefs at the altitude of 1,300 m. in its central part, whereas in the short and much narrower island of Rotti elevated reefs are known at an altitude of but 470 m. This will generally be true wherever the vertical motion prevails. The increase in breadth results from the fact that the vertical component of the rate of movement has generally been in the same direction near the coast as it has near the axis of the geanticline. If the distance of the geanticlinal axis from the coast line be considerable, the vertical component of the movement need not be the same for longitudinal and for transverse coasts. The length of the island may still increase though the breadth decreases, or both may decrease and the island get shorter and narrower, while the top is still moving upward. However, if the geanticline shows

a normal evolution, high reefs will always be found on large islands.

If this is not the case, and if adjacent small islands show elevated reefs at high altitudes, this points to the existence of fractures. This case is illustrated by the islands of the Babber group (Fig. 2). Some fine specimens of terraced islands are found in this group. In Babber the uppermost elevated reefs are found at an altitude of 650 m.;[1] the small island of Dai with a steep coast has fifteen terraces, the highest at 620 m. above sea-level; the small island of Dawera has probably sixteen terraces; and Daweloor has fourteen entirely covered with reefs. On Dawera the highest point is at an altitude of 328 m., and on Daweloor of 280 m. On Wetan, which also consists entirely of upheaved reefs, there are six or seven terraces in the southern part with a maximum altitude of 320 m. Wetan is separated from Babber by a narrow and deep strait without reefs.[2] Kisser, a small island of the Sermata group shows the same characteristics, having a fine terraced appearance with the highest reefs at an altitude of 147 m., though in the neighborhood of its coasts the sea bottom falls off rapidly to great depths.

FIG. 2.—The islands of the Babber group. (Southeastern Malay Archipelago.) Scale 1:3,000,000. 320, etc., altitude of the uppermost elevated reefs in meters. 200, 500, 1,000, submarine contours in meters.

THE EVOLUTION OF PARALLEL ROWS OF ISLANDS AND LONGITUDINAL FRACTURES STUDIED IN THE PROFILE

If we consider the evolution of geanticlines in a direction parallel to the geanticlinal axis, we find long and high islands where they are highest, and small and low islands at the depressions of the

[1] F. A. H. Weckherlin de Marez Oyens, "De Geologie van het Eiland Babber," *Handel. v. h. XIVe Nat. en Geneesk. Congres* 1913, pp. 463–68.

[2] R. D. M. Verbeek, *op. cit.*, p. 458.

axis. In the present-day stage of mountain-building this fact is illustrated by the Timor-Ceram row of islands, where a well-marked culmination occurs in the central part of Timor and well-marked depressions are found to the east and to the west of it. Secondary culminations and depressions are also found.

Sometimes two more or less parallel ranges of islands have the same direction as the geanticlinal axis. An example in the East Indian Archipelago is supplied by the islands of the Tenimber group, where the row which includes the main island Jamdena is

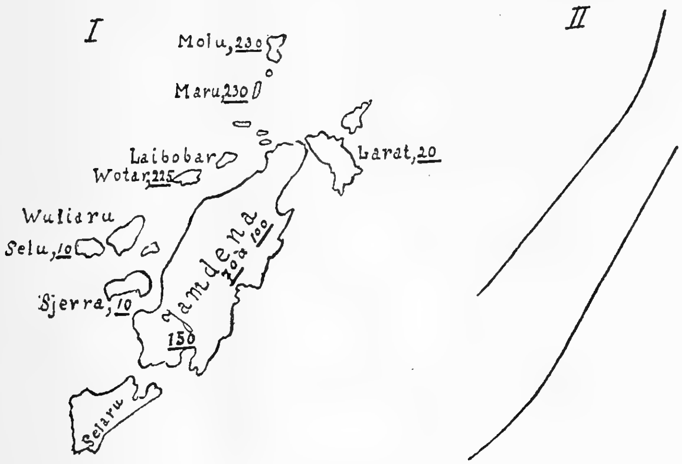

FIG. 3.—I, The islands of the Tenimber group. (Southeastern Malay Archipelago.) Scale 1:3,000,000. 225, etc., altitude of the uppermost elevated reefs in meters. II, The axes of the two secondary geanticlines (schematic representation).

accompanied by another row including the islands Selu, Wuliaru, Wotar, Laibobar, Maru, and Molu. The latter row differs from that of Jamdena in that it consists of smaller islands, although the elevated reefs are known at higher altitudes. On Wotar they are found at an altitude of 225 m., whereas on the main island Jamdena of the southern row the greatest height is at most 150 m. The reef cap, which covers Jamdena nearly continuously, is asymmetric, rising gradually from the northwestern coast in the direction of the main watershed of the island and thence descending rapidly toward the southeastern coast. I have explored portions of the coast of

the gently sloping northwestern part of this asymmetrical geanticline and found drowned river valleys which were observed far inland from the coast. Thus the upheaved island Jamdena is separated from the row of upheaved islands to the northwest by a zone which is covered by the sea, and in which during the youngest evolution of the geanticline positive movements have prevailed.

These facts can be explained in much the same way as has been done by Escher[1] for the group of islands southeast of Celebes which are known as the Tukang Besi Islands and which consist of four rows. Two of these rows consist of islands with elevated reefs which mark the anticlinal axes, whereas the two remaining show barrier reefs and atolls which mark the synclinal axes (Fig. 3).

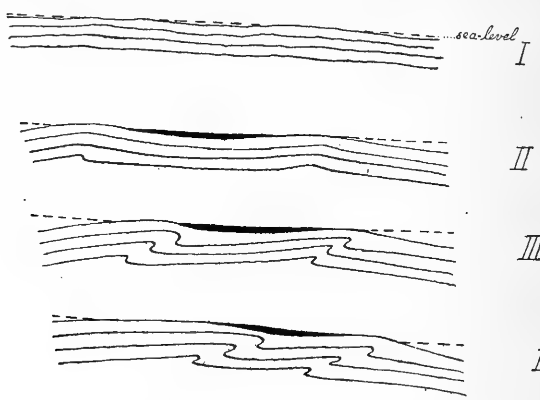

Fig. 4. One of the possible evolutions of two parallel rows of islands, of which different phases are represented in the southeastern Malay Archipelago (schematic representation). IV, Stage with elevated central basin.

We suppose that the geanticline at the Tenimber Islands is developing as two secondary geanticlines with an intermediate secondary geosyncline. The greater breadth of the islands in the southeastern secondary geanticline (although the reefs are not elevated to higher altitudes) may be the result of the prevalence of horizontal movements which caused the development of the asymmetrical reef cap (Fig. 4).

[1] B. G. Escher, "Atollen in den Nederlandsch Oost-Indischen Archipel: De Riffen in de Groep der Toekang Besi Eilanden," *Meded. Encyclop Bureau*, Afl. XXII (1920).

The further evolution of the geanticline can take place in different ways. If we suppose that in the next stage the plastic deformation of the northwestern secondary geanticline at greater depth causes chiefly a movement in a horizontal direction, the region of strongest upheaval will be displaced to the southeast. We may suppose that the rows of islands move in the direction of Australia, which for our considerations is the same as if Australia moved in the direction of the row of islands. Hobbs[1] has pointed out that mechanical difficulties disappear if the principal active forces involved in the folding of the Alps are considered as directed from the northwest toward the southeast. So far as our general conclusions are concerned, we may consider these movements as relative and not as absolute. The upper parts of the secondary geanticline do not move at the same rate and the higher parts of the folds were originally above the downward-moving secondary geosyncline. In a later stage of evolution these may be above the rising northwestern secondary geanticline and will be elevated above the sea.

Though differing in details, the geanticline of Timor may represent a later stage of geanticline evolution than the Tenimber Islands. In Pliocene time the geanticline near Timor was subjected to prolonged denudation and almost entirely disappeared below the sea. The crustal movements resulted in the development of two geanticlines and an intermediate, in part subdivided, geosyncline (cf. Molengraaff, *op. cit.*, p. 694), which became throughout fairly well filled by an accumulation of late Tertiary sediments deposited during a period of slow subsidence. Flexures and faults of considerable horizontal extent occur in the limbs of the geosyncline, which have caused the Pliocene strata within the basin to become bent abruptly upward near the edges. These longitudinal flexures and faults, which are essentially the same phenomenon, are the surface expression of an earlier, more plastic deformation at greater depth. Reefs and other littoral deposits spread over a great area, and after a certain period of evolution a great portion of Timor must have been covered by a sea full of

[1] W. H. Hobbs, "Mechanics of Formation of Arcuate Mountains," *Journal of Geology*, Vol. XXII (1914), p. 85.

coral islands and reefs, from which the islands emerged which are now the higher mountain groups of the present much enlarged island.

A similar stage of evolution is now to be observed in the same range of islands more to the east. The islands of the Sermata group clearly illustrate the movements of reefs in the period of development of the geanticline in which only its highest parts emerge from the sea as a group of smaller islands. The island of Luang has an altitude of 260 m. and, according to my observations, is built up entirely of Permian rocks. Together with two small islets at its southeastern extremity, it is fringed by a very broad reef, extending far to the east in the direction of Sermata and far to the west as well. Green islets far from the north coast, and barren, dry portions far from the south coast, mark the limits in northern and southern direction; beyond them the sea floor declines rapidly. Luang as well as the two small islets close to it rise up steeply from this broad reef, and no trace of elevated reefs was detected; the islands impress us as having originally formed one continuous whole and as having been separated by a positive movement, which may also account for the formation of the broad encircling reef.

In its eastern part the island of Moa consists of a low, very broad plateau of coral limestone, which rises scarcely more than 10–20 m. above the sea. From this plateau rises the steep Kerbau Mountain to an altitude of 400 m. Elevated reefs are lacking on the slopes of this mountain, and if the eastern part of Moa were a little lower, this region would present an aspect similar to that of Luang. The Island of Lakor, between Luang and Moa, consists of a low coral plateau, and Meaty Miarang forms the southern part of a large atolliform reef on the northern part of which lie the two low Ukenaö Islands. To the east of Luang and to the west of the eastern part of Moa the reefs are elevated to much greater altitudes and the group of the Sermata Islands shows a well-marked depression of the geanticlinal axis of the Timor-Ceram row. This part is much disturbed by transverse fractures and no sufficient data are available for judging whether the submersion observed on some islands is the consequence of the pitch of the geanticlinal axis only, or whether this region has passed, or will in the future

pass, through a stage with a secondary geosyncline between two secondary geanticlines.

After the Plio-Pleistocene reefs had been formed, a general elevation of the island of Timor took place. The elevation of the land has been somewhat greater at the edges of the secondary geosyncline than in the geosyncline itself, but the general movement resulted in the formation of a large anticline with the highest elevated reefs in the central part of the present island. In this latter stage of evolution the horizontal movements near the surface may have had a much smaller rate of movement than those at greater depth, while the central basin was gradually upheaved above the sea. The horizontal movement at greater depth may have prevailed in one of the secondary geanticlines only, but this is not a necessary condition. In our Figure 4 one of the possible modes of upheaval is represented.

DIFFERENT TYPES OF GEANTICLINAL MOVEMENT

The movement of a geanticline can be broadly described in the first place, in terms of the movements of the projections of the geanticlinal axis on the horizontal plane and on a vertical plane approximately parallel to the part of the geanticlinal axis under consideration. It is next of importance to take note of the movement of the section of the surface of the geanticline with a vertical plane at right angles to the geanticlinal axis. At the beginning of the movement we consider the geanticlinal axis to be a straight line; in a later stage this line will not be the geanticlinal axis, but for an approximate judgment this method is sufficient. The projections would undergo no changes in form if the geanticlinal axis was displaced parallel to itself. In general the vertical as well as the horizontal projection will develop a curved form. Some general types are given in Figure 5.

In Diagram I of Figure 5 the differences of plasticity and rate of movement between the surface and the deep-seated parts will have an influence on the development of longitudinal fractures only. The deformation of the sections perpendicular to the geanticlinal axis will be influenced by the place, the speed, and the duration of these fracture movements.

In Diagram II of the same figure the same considerations apply. The bending of a^1_2 will be much less than that in the figure, and the distinct traces of transverse fracture movements on the islands will disappear rapidly through erosion, although they may be perceptible near the transverse coasts.

In Diagram III of the figure more or less longitudinal fractures may develop, which in connection with the deformation of the

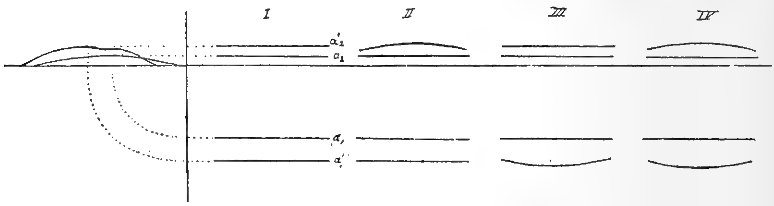

FIG. 5.—I, Displacement of the geanticlinal axis parallel to itself. II, III, and IV, Displacements in which the vertical or the horizontal projection, or both, have obtained a curved form.

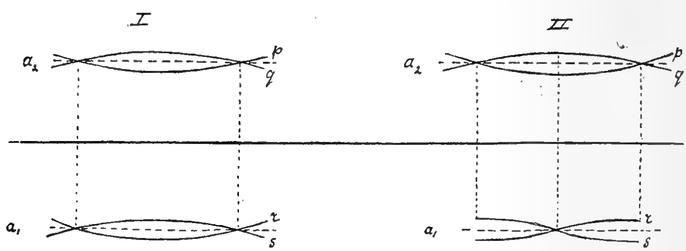

FIG. 6.—Deformations of the horizontal and vertical projections of the geanticlinal axis neglecting any displacements parallel to themselves.

sections perpendicular to the geanticlinal axis will be more or less important. Transverse fractures may be observable especially at the straits between the different islands of a row. Diagram IV is a combination of Diagrams II and III.

If in Diagram II, a^1_2 has one or more bending-points, which is equivalent to the development of transverse folds normal to the geanticlinal axis, then the place, rate, and duration of the transverse fracture movements near the surface may be strongly influenced by these folds. The same considerations are applicable to IIIa^1_1, and to IVa^1_2 or IVa^1_1, or to both of them.

If we consider Diagram IV, the more general type of deformation, supposing that the horizontal and the vertical projections of the geanticlinal axis have an equal number of bending-points, then two different types can be distinguished according as the bending-points of the horizontal and vertical projections alternate or do not (Fig. 6).

In this figure the combinations *p-r*, *q-s*, *q-r*, and *p-s* are different curves in space to which the originally rectilinear geanticlinal axis *a* has been distorted. The displacement of the geanticlinal axis parallel to itself and the distortion of the sections perpendicular to the geanticlinal axis are left out of consideration. The bending of *p* and *q* is much less than that which is shown in the figures. It would be more important if a strong compression had been acting in the direction of the geanticlinal axis from which would result a deformation to transverse folds normal to the geanticlinal axis. In this connection it is necessary to consider the geanticline over sufficiently long distances to obtain a judgment concerning the deformation of the vertical projection of the geanticlinal axis.

APPLICATION TO THE TRANSVERSE FRACTURES OF THE TIMOR-CERAM ROW OF ISLANDS

If considered over large distances it might seem that the geanticline, Sumba-Rotti-Timor-Sermata Islands, represents approximately Type I. The uppermost elevated reefs are found in Central Timor at an altitude of 1,300 m.; in West Timor, southeast of Kupang, they are at a height of 500 m., on Rotti at 470 m., and on Savu at 300 m. In East Timor the altitude is estimated at 600 m., on the islands farther to the east such reefs are known at altitudes of 140 m. on Letti and of 20 m. on Lakor, while on Luang no elevated reefs are found (cf. also Fig. 7). Thus the part, Sumba-Savu-Rotti-West Timor of the geanticline, would represent approximately I *p-r*, and the part, Central and East Timor–islands farther to the east, would represent I *q-s*.

If considered in detail the deformation is much more complicated. The deformation of the vertical projection is very slight and in many cases it is not exactly known. If the motion of the geanticlinal axis parallel to itself be neglected, this projection

nearly coincides with a, and the distinction between I a_2-r, I a_2-s and II a_2-r, II a_2-s disappears. Between Rotti and West Timor (Fig. 7) II p-r may be represented, but the deformation of the horizontal projection is the only important one.

The strait between Timor and Rotti coincides with a bending-point of the horizontal projection of the geanticlinal axis. In seeking an explanation for the existence of this strait, we might suggest the pitch of the geanticlinal axis on both sides of the strait, while at the place of the strait the axis could disappear below sea-level. But if considered in detail, this explanation alone is not applicable. On Rotti we find between the main island and the peninsula of Landu a narrow strait which only recently has been

FIG. 7.—I, Rotti, Timor, and the Sermata Islands. 470, etc., altitude of the uppermost reefs in meters. Straits and transverse dislocations are near the bending-points of II. II, Geanticlinal axis with bending-points between Timor and Rotti and between Timor and the Sermata Islands.

filled up by a mud bank still inundated at spring tide. At both sides of the narrow strait high walls of elevated coral limestone occur, and during an exploration of the island I found a small isolated rock composed of coral limestone which emerges from the mud in the middle of the strait. These facts point to the existence of transverse gaping fractures formed by a movement with a component normal to the fracture plane. We have already mentioned similar facts in connection with the groups of small islands having high reefs.

Another example of the same sort is found to the east of Timor (Fig. 7). Considering the large bendings only, a bending-point is located between East Timor and the Babber group; but if considered in detail bendings of smaller amount may also be observed. We note a bending of the geanticlinal axis between East Timor and

Letti of the Sermata group, and in the neighborhood of the bending-point we observe the northern, non-harmonic position of the small island Kisser which is covered by elevated reefs and surrounded by deep seas. There is here again the evidence that bending of the geanticline at greater depth is accompanied by transverse fractures near the surface. The fractures which occur farther to the east and their connection with the sharp bending in the 200 m. contour line of the Sahul shelf has already been discussed in earlier papers.[1]

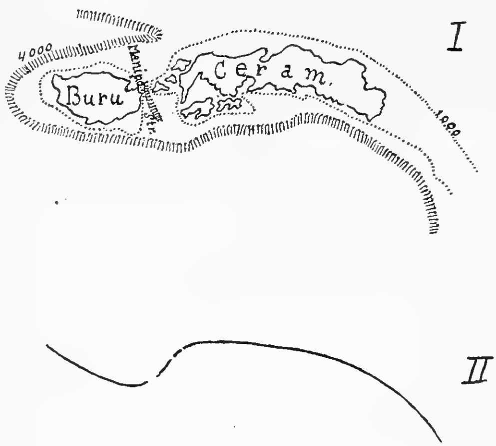

FIG. 8.—I, The deep Manipa Strait (+4800 m.) between Ceram and Buru. 1000, 4000, submarine contours. II, Geanticlinal axis with strong transverse dislocations near the bending-point.

Still another example is found between Ceram and Buru (Fig. 8). A very striking irregularity in this portion of the geanticline is the narrow Manipa Strait, nearly 5,000 m. deep between Ceram and Buru, here also near the bending-point of the horizontal projection of the geanticlinal axis. If this bending-point is not so clearly visible in the present topography, for the reason that the fracture movements are very strong, it may be inferred from the strike of the Tertiary mountain range. In West Buru and in the greater part of Ceram this strike is about NW.–SE., whereas in West Ceram and in the islands between Ceram and Buru it is E. NE., and NE. strikes also have been observed (cf. Fig. 1).[2] Thus the Tertiary mountain range displays a considerable bending from Ceram to Buru. As we have pointed out, the strike of the folds and the overthrusts of the Tertiary phase of crustal movement, and

[1] H. A. Brouwer, *loc. cit.*

[2] L. Rutten and W. Hotz, "De geologische Expeditie naar Ceram," *Tydschr. Kon. Ned. Aardr. Gen.*, Vol. XXXVI (1919), 9e Verslag.

the fractured and faulted crust near the surface of the youngest phase of the deformation are the result of different stages of the same process. As has been stated above, p and q will nearly coincide with a_2, and the bending of p and q is much less than is represented in the figures. In the case of the Strait of Manipa a compression may even have been acting in the direction of the geanticlinal axis so that the origin of the strait may in part be due to transverse folding. But no sufficient data are as yet available for an exact judgment on the problem of deformation in space.

We have seen that in the large bendings of the geanticlinal axis a distinction between Types I and II can be made, though if considered for bending at relatively small distances, these two types are very similar for the reason that p and q nearly coincide with a_2. In the Timor-Ceram row the following rule seems to be approximately applicable:

Considerable transverse fractures near the surface of the moving geanticline coincide with bending-points of the horizontal projection of the geanticlinal axis.

In most cases it is clearly observable that the fractures near the surface have been formed by a movement having an important component normal to the fracture plane, and that the fractures near the bending-points are the surface expression of *differences in rate of movement* of neighboring points in the horizontal projection of the geanticlinal axis.

RELATIVE AND ABSOLUTE HORIZONTAL MOVEMENT

If we neglect the displacement of the geanticlinal axis parallel to itself, as has been done, we find evidence only for the relative horizontal displacements of different points in the geanticlinal axis. The absolute horizontal movement may be considerable, but it cannot be inferred from the surface characters of the present geanticlines. If our interpretation of the evolution of the central basin of Timor is correct, important absolute horizontal movements must have taken place at greater depth, while the superficial parts moved at a slower rate. This conception agrees with the interpretation of the evolution of the Western Alps, as this has been

demonstrated by Argand.[1] Here likewise we see that in Mesozoic time geanticlines formed, separated by geosynclines, and that these have been moved in a horizontal direction. It may be that the southeastern Indian Archipelago will in the future arrive at the same stage as was long before reached in the Alps. As the horizontal movements proceed, the sea basins will narrow, and eventually the masses of the deeper parts of the present rows of islands will be pushed over the present Australian continent and the Sahul shelf which extends its borders. For a judgment, whether the active force tending to produce movement is directed to the southeast or to the northwest, as would follow from the conceptions of Hobbs[2] and Wegener,[3] no sufficient data are available.

[1] E. Argand, "Sur l'arc des Alpes occidencales," *Eclogae Geol. Helv.*, Vol. XIV (1916), p. 179.

[2] W. H. Hobbs, *op. cit.*, p. 91.

[3] A. Wegener, "Die Entstehung der Kontinente und Ozeane," *Die Wissenschaft*, 1920.

Reviews

The Oil and Gas Resources of Kentucky. By Willard Rouse Jillson, Kentucky State Geologist. Department of Geology and Forestry, Frankfort, Ky. 1919. Pp. 630.

This volume is a review and summary of the known oil and gas resources of Kentucky, found in strata ranging from the Ordovician to the Pennsylvanian. The "Calciferous" (Lower Ordovician) contains a small amount of oil and gas, but it is doubtful if it will ever be important commercially. The "Trenton" formation (Middle Ordovician) is an important source of oil, and the Cincinnatian has yielded some. The "Niagaran" (Silurian) is productive, but the Onondagan formation ("Corniferous," Middle Devonian) holds first place in the production of petroleum and natural gas in the state. The "Black Shale" (Upper Devonian) does not yield petroleum in commercial quantities, but some gas has been derived from it. The Waverly series (Lower Mississippian) contains oil and gas sands of importance. The St. Genevieve (=St. Louis limestone, Middle Mississippian) has not yielded petroleum on a commercial scale, but has yielded much gas. The Chester or Mauch Chunk formation (Upper Mississippian) is a good producer in eastern Kentucky. Both oil and gas occur in the Pottsville Conglomerate (Pennsylvanian). Neither petroleum nor gas is known to exist in formations younger than the Pennsylvanian, within the state.

R. A. J.

Descriptive Mineralogy. By W. S. Bayley. Appleton & Co., 1917. Pp. 542+xvii. $3.50.

To quote from the Preface, "The following pages are presented with the purpose of affording students a comprehensive view of modern mineralogy rather than a detailed knowledge of many minerals. . . . . The volume is not a reference book. It is offered solely as a textbook." Bearing these statements in mind, the work is one which may be highly commended, especially since it does not invade an overcrowded field.

As the title indicates, it is almost purely a descriptive mineralogy, lacking a discussion of crystallography, but containing material on the

composition, classification, formation, and alteration of minerals, as well as the principles and methods of blowpipe analysis. Appendices contain lists of minerals arranged according to their principal constituents and to their mode of crystallization, a list of reference books, and an abbreviated "key to the determination of minerals." The key is a device which classifies minerals according to luster, streak, color, and hardness and gives merely the pages in the main part of the text, where the detailed descriptions of the minerals may be found.

Perhaps the most noticeable defect is the paucity of photographs (less than forty), which probably accounts for the low price of the work. However, there are numerous drawings which remedy this deficiency to a large extent.

Minerals are classified according to their chemical composition. The arrangement of the silicates, a most difficult problem, is especially worthy of favorable comment. The book should be of great value as a text for advanced work in descriptive mineralogy.

D. J. F.

Detailed Report on Webster County. By D. B. Reger, West Virginia Geological Survey, Morgantown, W.Va. 1920. 671+xvi pages, 35 halftone plates, and 24 zinc etchings in the text, accompanied by a separate case of topographic and geologic maps. Price, including case of maps, charges prepaid, $3.00. Extra copies of topographic map, 75 cents, of the geologic map, $1.00.

Webster County contains the northward extension of the famous New River Coal group, as also the Kanawha group and the lower members of the Allegheny Series in its northern portion.

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Geologic Reconnaisance in Baja California. By N. H. Darton.

Pennsylvanian Stratigraphy of North Central Texas. By Raymond C. Moore and Frederick B. Plummer.

The Relation of the Physical Properties of Natural Glasses to Their Chemical Composition. By William O. George.

Pleistocene Mollusca from Northwestern and Central Illinois. By Frank Collins Baker.

The Character of the Stratification of the Sediments in the Recent Delta of Fraser River, British Columbia, Canada. By W. A. Johnston.

Notes on a Recent Collection of Fossil Plants from the Dalles Group of Oregon. By Ralph W. Chaney.

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THE MARINE TERTIARY OF THE WEST COAST OF THE UNITED STATES: ITS SEQUENCE, PALEOGEOGRAPHY, AND THE PROBLEMS OF CORRELATION[1]

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INTRODUCTION

Considerable work has been done on the marine Tertiary deposits of the West Coast during the past ten years, and some of the discoveries that have been made have greatly modified many of our previous conceptions. The purpose of this paper is to review some of the most salient facts concerning the stratigraphic divisions, paleogeography, and correlation of these horizons in order to give the reader some idea of the present status of this knowledge.

The paper includes a correlation table of the marine West Coast Tertiary. The construction of such a table is a very difficult task, and it will undoubtedly be a good many years before a table can be made which will be satisfactory to everyone working in this field. None of the West Coast Tertiary horizons has been thoroughly studied: there is a notable lack of detailed mapping, and most of the faunas have been inadequately monographed. The West Coast Tertiary still offers some of the most important problems for stratigraphic and paleontological research in the United States, and if the reader can obtain from this paper some

[1] Read before Geological Society of America, December, 1920.

idea of the problems involved, the task will have been well worth while.

There are several factors, aside from the lack of a sufficient number of trained workers, that have hindered the progress of correlation of West Coast horizons. These factors may be considered under the following headings: (1) temperature differentiation (2) geographical isolation, and (3) poor preservation.

1. It is well known that the marine faunas living on the Pacific Coast can be separated into distinct faunal and geographical provinces. In this respect the West Coast of North America is typical of the whole Pacific border. For example, the fauna found off the coast of Panama is very different from that living along the coast of southern California, and the latter has very little in common with that off the coast of Alaska, while faunas from some of the intermediate areas are almost equally distinct.[1] It is generally recognized that Pleistocene and Recent times mark one of the maximum periods of emergence of all the continents. While this is not true in so great a measure of all the periods of the Tertiary, it is well known that the North American continent was submerged only on its borders, and that during a large part of this time the Pacific and Atlantic oceans were disconnected. The study of the Tertiary faunas along the coast discloses marked evidences of temperature differentiation; the faunas of the north having a more boreal aspect than those of the south.[2] This differentiation was most exterme in the Pliocene and Upper Miocene, and it is undoubtedly because of this that there has been so much confusion in the past in the correlation of the deposits from various sections along the coast now referred to those horizons. There is good evidence of temperature differentiation during the Oligocene and Middle Miocene, and what is more interesting is that accumulated evidence seems to show that this differentiation of the faunas had its effects even as far back as Eocene times.

[1] W. D. Dall, *Summary of the Marne Shellbearing Mollusks of the Northwest Coast of America, Bulletin 112, United States Natural Museum* (1921), pp. 1–213.

[2] J. P. Smith, "Climatic Relations of the Tertiary and Quaternary Faunas of the California Region," *Proc. Cal. Acad. Sci.*, Fourth Series, Vol. IX (1919), No. 4, pp. 123–73.

2. The second factor, that of geographical isolation, was very probably an important one, and if so was the result of numerous partially isolated local basins of deposition. The sediments of the Tertiary were for the most part laid down in geosynclinal troughs which paralleled the present Coast ranges. The number and position of these troughs has varied through the different periods and epochs of deposition. There was therefore a condition similar to that which existed in the Appalachian geosyncline during the Paleozoic. Great thicknesses of clastic sediments, in aggregate exceeding 40,000 feet, were deposited on the West Coast during Tertiary time. These Tertiary basins existed either as large embayments or long inland seas, some of the latter of which were comparable in size to the Mediterranean and were probably nearly as well separated from the main ocean basin. These conditions produced marked local environments, with corresponding local changes in the faunas. It is very probable that the faunas in each basin derived certain peculiar characteristics due to isolation alone.

3. Still another factor that has brought about difficulties in correlation in the West Coast Tertiary has been the rather general poor preservation of the fossil material. The Tertiary beds have been extensively folded and tilted, and this deformation has resulted in the leaching of the original material of the shells, especially in the sandstones and shales. Intensive collecting will in time remedy this difficulty as well as bring to our knowledge a larger number of localities where the fossils are in a better state of preservation.

In presenting a correlation table of this kind, one of the first things that will be asked is the author's point of view in attacking the problems. The point of view accepted is that diastrophism is the fundamental basis for differentiating geologic divisions. In other words, the divisions recognized in this paper have been made on the basis of stratigraphic breaks which are believed to be more than local. It is important to note that every stratigraphic unit thus recognized is also represented by a distinctive fauna.

The paleogeographic maps presented in this paper are not accurate in detail and will undoubtedly be modified by future work. The present knowledge of the geology of the Coast ranges

does not enable one to show the exact location of the shore lines of all the seas that have occupied this general area. However, it is believed that the plates show the approximate distribution of the seas and will give the reader some conception of the location of the most important land masses, the degree of isolation of the basins, and the present known extent of the Tertiary horizons in California.

TERTIARY DIVISIONS

There are at least five major divisions of the Tertiary of the West Coast which in the writer's estimation might be recognized as representing true periods. Each one of these five major divisions is composed of more than one epoch of deposition. Each epoch is represented by distinct faunas which lived in distinct seas. The deposits belonging to each of the major divisions will be referred to as a "series"; thus, the Eocene, Oligocene, Lower-Middle Miocene, Upper Miocene, and Pliocene series. To the deposits of each epoch of deposition the term "group" has been applied. The term "formation" is reserved for the lithologic member within the group.

EOCENE

During the Eocene period of the West Coast there were at least three epochs of deposition, as indicated in the correlation table, and there is a suggestion that there were four and possibly five. At the present time, however, only three distinct stratigraphic divisions have been definitely separated. These are the Martinez (Lower Eocene), Meganos (Middle Eocene), and the Tejon (Upper Eocene). Crustal movements of considerable magnitude separated these epochs. In the region of Mount Diablo, middle California, there is a difference in dip and strike between the Martinez and Meganos groups. Over large areas along the coast where beds of the Meganos and Tejon deposits occur in contact, there is an angular unconformity separating the two. The marine faunas of these three divisions of the Eocene differ greatly from each other, further substantiating the stratigraphic evidence of marked

MARINE TERTIARY OF THE WEST COAST

CONTINENTAL DEPOSITS

OTHER SECTIONS
According to T. W. Vaughan

Mount Diablo Region

Los Angeles to Santa Ana Mountains

Generalized Section of Oregon, Washington

Pliocene

Upper Miocene

Lower Miocene

Oligocene

Eocene

Santa Margarita

Temblor
"Monterey Shale"
Pecten andersoni Arnold
Turritella ocoyana Conrad
Pecten peckhami Gabb

Temblor
"Salinas Shale"
Pecten andersoni Arnold
Pecten peckhami Gabb
Turritella ocoyana Conrad

Temblor
"Big Blue"
Merychippus calamarius
"Vaqueros"
Pecten andersoni Arnold
Turritella ocoyana Conrad
Scutella merriami Pack

Temblor
(Sandstones and Shales)
Pecten andersoni Arnold
Pecten peckhami Gabb
Chione temblorensis Anderson
Turritella ocoyana Conrad

Temblor
Turritella ocoyana Conrad

"Modelo"

Temblor
Sandstone and Shale

Temblor
Turritella ocoyana Conrad

Temblor
"Wahkiakum"
Turritella ocoyana Conrad

Called "Maricopa Shales" by Pack

5000' organic shale

6000' organic shales

"SALINAS SHALE"

4500' Organic Shale

"PUENTE SANDSTONE"
(Including Vaqueros)

Vaqueros
Turritella inezana Conrad

Vaqueros
Turritella inezana Conrad
Pecten magnolia Conrad
Pecten perrini Arnold

Vaqueros
Pecten perrini Arnold
Turritella inezana Conrad

Vaqueros
Pecten sespeensis Arnold
Pecten sespeensis Arnold var. hydei
Pecten magnolia Conrad
Turritella inezana Conrad
Scutella fairbanksi Arnold

Vaqueros
Pecten sespeensis Arnold
Pecten sespeensis Arnold var. hydei
Pecten magnolia Conrad
Turritella inezana Conrad
Scutella fairbanksi Arnold

Vaqueros
Turritella inezana Conrad

(Including Oligocene)

San Lorenzo

"Sespe"

Sespe

Tortonian

Helvetian

Burdigalian

Aquitanian

Rupelian

Chattahoochee

hiatuses.[1] Thus, on the West Coast, the Eocene period may be definitely stated to be made up of at least three epochs of deposition and should be recognized as a true period rather than as an epoch.

Correlation of Eocene deposits.—The evidence for the correlation of the West Coast marine Eocene with that of the Gulf and East Coast provinces and through them with Europe is based upon the identity of species or the presence of closely related forms common to the two regions. The evidence appears to be much better for the correlation of the Meganos and the Tejon (Middle and Upper Eocene) than for the Lower or Martinez group. There can be little doubt but that during those epochs of time there was a direct connection between the Gulf of Mexico and the Pacific Ocean.

Climate.—The climate during the Eocene of the West Coast was subtropical or possibly warm temperate rather than tropical. The arkosic character of the Meganos deposits, a character very general on the West Coast, strongly suggests that we are dealing with deposits which were derived from an arid coast, while Tejon sandstones, at most localities in California, are composed almost entirely of pure quartz grains, indicating humid climatic conditions at that time.[2]

(Fig. 1.) *Paleogeography.*—As indicated by the paleogeographic maps, the deposits of the Martinez group (Lower Eocene, Fig. 2) were laid down in much more limited basins than those of the Meganos and Tejon groups. Apparently there were at least four separate basins in California, the connections between which were indirect. It seems very probable that when the faunas obtained from these four areas have been more fully described, we shall find that the geographical factor has caused considerable difference between them.

[1] R. E. Dickerson, "Fauna of the Martinez Eocene of California," *Bull. Dept. Geol., Univ. Cal.*, Vol. VIII (1914), No. 6, pp. 61–180. B. L. Clark, "The Meganos Group, a Newly Recognized Division in the Eocene of California," *Bull. Geol. Soc. Am.*, Vol. XXVIII (1918), pp. 218–96; "Stratigraphy and Faunal Relationships of the Meganos Group, Middle Eocene of California," *Jour. Geol.*, Vol. XXIX (1921), No. 2, pp. 125–65.

[2] R. E. Dickerson, "Climatic Zones of Martinez Eocene Time," *Proc. Cal. Acad. Sci.*, Fourth Series, Vol. VII (1917), No. 7, pp. 193–96.

The Meganos and the Tejon seas (Figs. 3 and 4) were somewhat similar in outline. In middle California the deposits of these epochs were laid down in a great trough of which the present Great Valley of California is a remnant. The Meganos sea was the wider of the two Eocene seas that occupied this depression

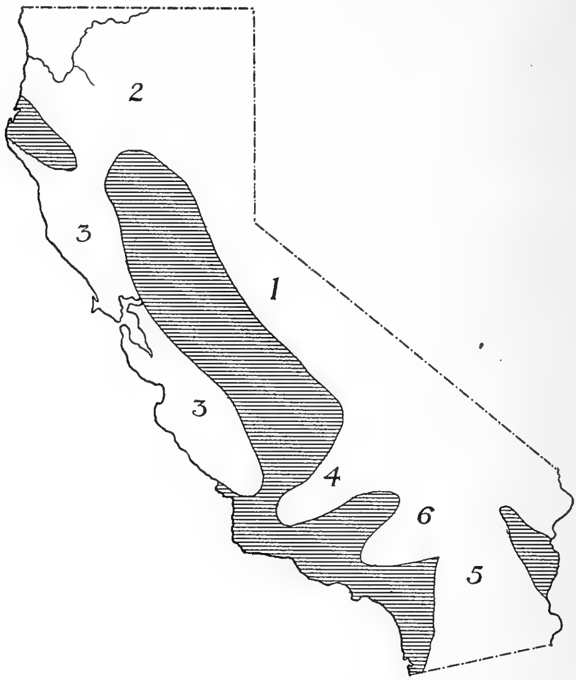

FIG. 1.—A key map showing the general distribution of positive and negative areas in California during the Tertiary. Of all the positive areas outlined, No. 6, the Santa Monica Mountain area, is the most problematical. (1) Sierra Nevada area; (2) Klamath Mountain area; (3) Coast Range area; (4) Tehachapi Peninsula; (5) Sierra Madre, San Bernardino, San Jacinto Mountain area; (6) Santa Monica Mountain area.

and was connected with an east and west trough in southern California in the region of the present Santa Ynez Mountains. These two general areas of deposition existed throughout the Tertiary. They were bordered by areas or zones of uplift which have also been more or less permanent. East of the great north and south trough was the Sierra Nevada block which dates back to the Upper

Jurassic. To the west there was a positive area covering the present western side of the Coast ranges of middle California. Apparently throughout the entire Eocene period this area to the west was a positive block; the absence of either continental or marine deposits in this area is the chief basis for the conclusion.

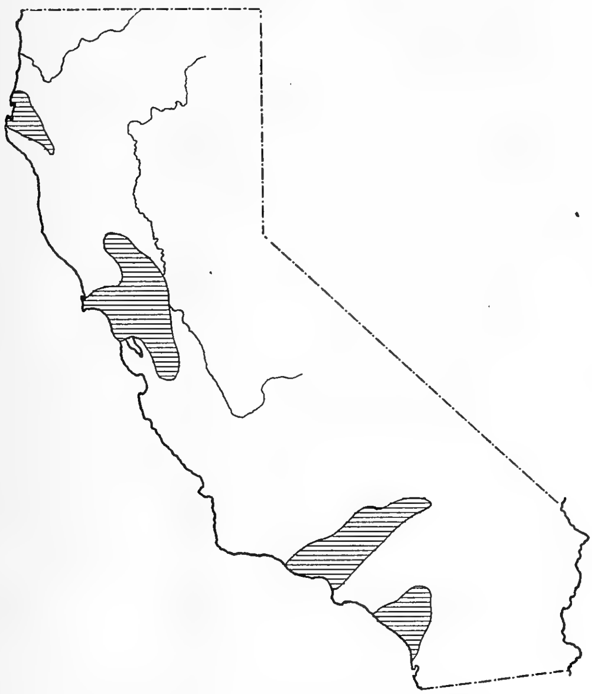

FIG. 2.—Martinez (Lower Eocene)

The smaller positive and negative areas which existed during the Oligocene, Miocene, and Pliocene in this western area were not differentiated during the Eocene.

One of the most permanent positive areas of the Tertiary was that which existed in the region now occupied by the Tehachapi and San Emigdio Mountains (Fig. 1). This area formed the east-west peninsula which separated the northern from the southern basins. The extent of this peninsula varied considerably during the different epochs of deposition. The area now covered by the

Santa Monica Mountains was apparently part of an early positive block bordering the east–west trough mentioned above. To the north of this trough the region now occupied by the Santa Ynez Mountains constituted the westward extension of the old Tehachapi peninsula.

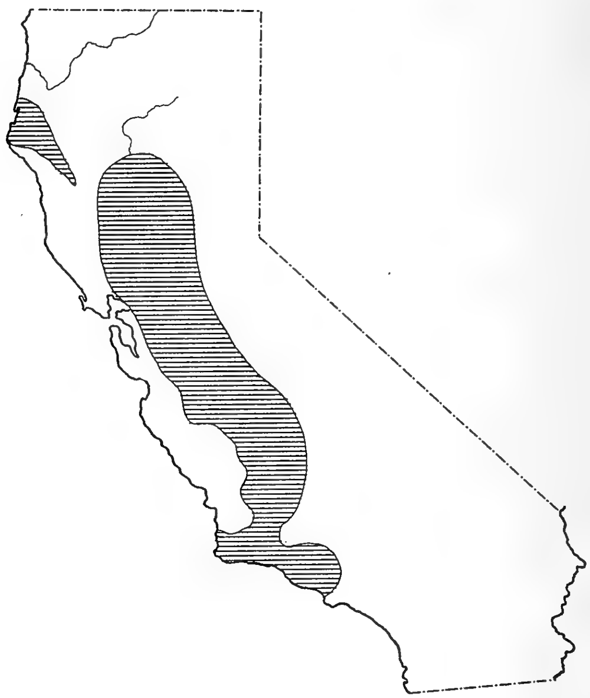

FIG. 3.—Meganos (Middle Eocene)

Some of these old positive areas were so persistent throughout the Tertiary time that they might well be given names, following, on a smaller scale, the example of Schuchert in his *Paleogeography of North America*. Certain of these positive and negative areas have been persistent throughout all Tertiary time, while others were formed at a later date. Also, it is worthy of note that the old positive and negative areas had the same trend as the present mountain ranges.

OLIGOCENE

Accumulated evidence appears to show that there were at least two distinct epochs of deposition on the West Coast during the Oligocene. The two epochs are represented by the Molopophorus lincolnensis and Acila gettysburgensis zones of Dr. C. E. Weaver. They

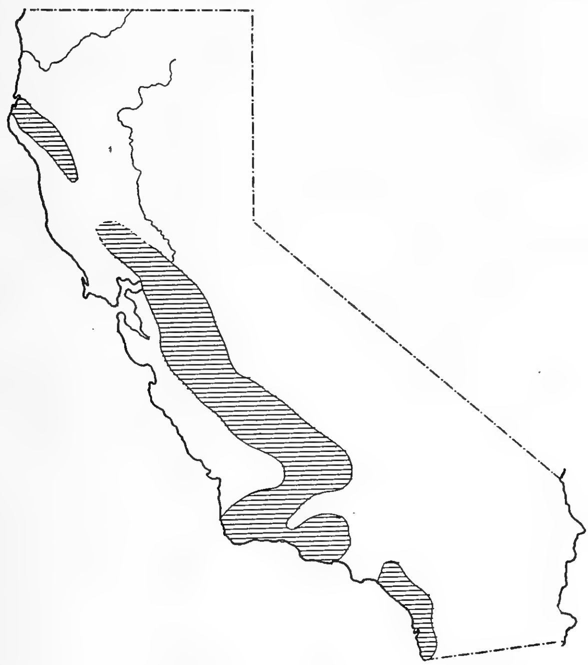

FIG. 4.—Tejon (Upper Eocene)

will be referred to as the Lincoln and San Lorenzo.[1] Recent field work of the writer has shown fairly conclusively that there were at least two distinct epochs of deposition in Oregon and Washington. He believes that this will prove to be the case in California when more detailed work on the stratigraphy of the Oligocene over wider areas has been completed. It is certain that the two faunas first

[1] C. E. Weaver, "Tertiary Formations of Western Washington," *Wash. Geol. Surv., Bull. 13* (1916), pp. 1–21; "Preliminary Note on the Paleontology of Western Washington," *Wash. Geol. Surv. Bull., 15* (1912), pp. 1–80; "Tertiary Faunal Horizons of Western Washington," *Pub. Univ. Wash.*, Vol. I (1916), No. 1, pp. 1–67.

differentiated in Washington are also represented in the same sequence in California, and tentatively we may consider the Oligocene series as being made up of two distinct parts, referred to in the correlation table as the San Lorenzo series.

The aggregate thickness of the marine beds of the Upper and Lower Oligocene of the West Coast exceeds 10,000 feet. A large part of these sediments consists of shales and shaly sandstones.

Correlation.—The evidence for the correlation of the West Coast marine Oligocene deposits is indirect. No molluscan species or even apparently related forms have been recognized as common to the Oligocene of the West Coast and the Gulf province. The faunal evidence at hand seems to show that after the close of the Tejon epoch (Upper Eocene) there was no direct connection between the Atlantic and the Pacific Coast basins.

Dr. Ralph Arnold was the first to announce the presence of Oligocene in California. The type section of the San Lorenzo is in the Santa Cruz Mountains of the Santa Cruz Quadrangle, California. Dr. Arnold concluded that this formation is of Oligocene age because of its stratigraphic position between beds generally recognized as belonging to the Upper Eocene and Lower Miocene (Vaqueros) age. He observed that the fauna of the San Lorenzo appeared to have both Eocene and Miocene affinities.[1] Later studies of the faunas of the Lincoln and San Lorenzo horizons have borne out Arnold's original conclusions.[2] At the time Arnold did his work the Lincoln horizon had not been differentiated. The fauna of this horizon shows a much closer relationship to that of the Tejon (Upper Eocene) than to that of the Lower Miocene, while the fauna of the San Lorenzo horizon, equivalent to Weaver's Acila gettysburgensis zone, has a Miocene aspect, a fairly large number of the genera and species being common to the two.

[1] R. Arnold, "Tertiary and Quaternary Pectens of California," *U.S. Geol. Surv., Prof. Paper 47* (1906). J. C. Branner, F. G. Newsom, and R. Arnold, *U.S. Geol. Surv., Folio 163*, Santa Cruz Folio.

[2] B. L. Clark, "Occurrence of Oligocene in the Contra Costa Hills of Middle California," *Bull. Dept. Geol., Univ. Cal.*, Vol. IX (1915), No. 2, pp. 9–21; "San Lorenzo Series of Middle California," *Bull. Dept. Geol., Univ. Cal.*, Vol. XI (1918), No. 2, pp. 45–234. B. L. Clark and R. Arnold, "Marine Oligocene of the West Coast of North America," *Bull. Geol. Soc. Amer.*, Vol. XXIX (1918), pp. 297–308.

Climate.—The temperature conditions during the Oligocene time were fairly uniform along the West Coast as far north as Alaska. The waters of the Lower Oligocene, judging from the molluscan fauna, were subtropical to warm-temperate, while those of the Upper Oligocene sea were more temperate.[1] Thus the fauna of the San Lorenzo horizon (Upper Oligocene) is more closely related to that living off the coast of California, Oregon, and Washington at the present time than it is to that of the Lincoln.

Paleogeography.—The distribution of the Lower Oligocene deposits (the Lincoln horizon) (Fig. 5) corresponds closely to that of the Tejon (Upper Eocene). In California there was a long inland trough corresponding closely to, though somewhat wider than, the present Great Valley of California. The presence of great thicknesses of organic shales of the Kreyenhagen formation, from which the oil of the Coalinga field is derived, indicates that the deepest portion of the Lower Oligocene trough was along the western border of the present San Joaquin Valley.[2]

The distribution of the Upper Oligocene (San Lorenzo) is very different from that of the Lower Oligocene. In California there were two limited basins of deposition, one in middle California in the vicinity of San Francisco, and one in the region of the southern end of the San Joaquin Valley.

A fauna referred to the Oligocene which may be Upper Eocene.—The fauna of the Tejon (Upper Eocene) of the West Coast, as has already been stated, can probably be correlated with the upper Claiborne horizon of the Eocene of the Gulf province, but whether or not there is a fauna on the West Coast that is equivalent to the Jackson horizon of that province can only be proved by further detailed study.

The possibility of referring the Molopophorus lincolnensis zone, now considered Lower Oligocene, to the Jackson stage has been considered. Dr. C. E. Weaver has listed a number of species

[1] R. E. Dickerson, "Climate and Its Influence upon the Oligocene Faunas of the Pacific Coast," *Proc. Cal. Acad. Sci.*, Fourth Series, Vol. VII (1917), No. 6, pp. 157–92.

[2] R. W. Anderson and R. W. Pack, "Geology and Oil Resources of the West Border of the San Joaquin Valley North of Coalinga, California," *U.S. Geol. Surv. Bull. 603* (1915), pp. 74–78.

in the former horizon that are also found in the Tejon and the generic assemblages are notably similar. The Molopophorus lincolnensis zone, however, shows a closer relationship to the Acila gettysburgensis zone (Upper Oligocene) than to the Tejon, and subsequent work on the fauna of the Molopophorus lincolnensis

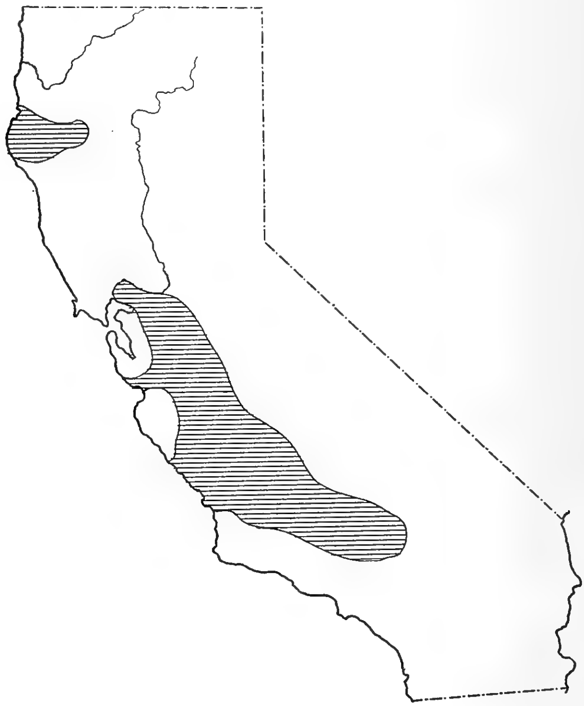

FIG. 5.—Lower Oligocene

zone has shown that there are fewer species common to the Tejon than Dr. Weaver supposed.[1]

There is still another fauna which may represent an Eocene stage higher than the Tejon. During the year 1912 a collection was made by Mr. F. M. Anderson and Mr. Bruce Martin, at that time curator and assistant curator in the department of paleontology of the California Academy of Sciences, from the Greise Ranch

[1] C. E. Weaver, "Preliminary Report on the Tertiary Paleontology of Western Washington," *Wash. Geol. Surv., Bull. 15* (1912), p. 16; "Tertiary Formations of Western Washington," *Wash. Geol. Surv., Bull. 13* (1916), p. 167.

near the town of Vader in southern Washington. This collection came from beds unconformable on the Tejon, and below strata containing a typical Molopophorus lincolnensis fauna. The fauna was described by Dr. R. E. Dickerson and consisted of forty-eight species, of which thirty-six were considered new and thirteen were determined as common to the Molopophorus fauna.[1] The writer has had the opportunity of making larger collections from the Greise Ranch locality, and with Dr. G. D. Hanna, present curator of the department of paleontology of the California Academy of Sciences, has re-worked the fauna listed and described by Dr. Dickerson. The results of this work show quite conclusively that there is very little, if anything, in common between this fauna and that of the Molopophorus lincolnensis zone. The fauna at the present time consists of about seventy-five species and is very distinct from any other known fauna on the Pacific Coast. None of these species have been definitely determined as common to either the Tejon or the Molopophorus lincolnensis zone, and the stratigraphic position of the horizon renders it possible that these beds are equivalent to the Jackson of the Gulf province.

MIOCENE

The marine Miocene of the West Coast is divisible, both on the basis of stratigraphy and fauna, into two major series each of which contains minor horizons.

The portion of the geological section referable to the Monterey series (Lower-Middle Miocene) contains two fairly distinct faunas and two epochs of deposition, at least in certain areas of the state of California. The lower of these two divisions is the Vaqueros group, sometimes referred to as the "Turritella inezana" zone. The upper division of the Monterey series is herein referred to as the Temblor group and is represented by the fauna of the "Turritella ocoyana" zone. The deposits of the Vaqueros Sea covered a much more limited area than those of the Temblor, and have not been found in Oregon or Washington.

[1] R. E. Dickerson, "Climate and Its Influence upon the Oligocene Faunas of the Pacific Coast with Descriptions of Some New Species from the Molopophorus Lincolnensis Zone." *Proc. Cal. Acad. Sci.*, Fourth Series, Vol. VII (1917), No. 6, pp. 157–92.

The upper major Miocene division constitutes the San Pablo series, which is composed of three minor stratigraphic and faunal divisions, the Briones, Cierbo, and Santa Margarita groups. As will be brought out later, each one of these groups represents a distinct sequence of deposition and possesses a fairly distinctive fauna.

Monterey series.—Whether or not the Vaqueros and Temblor represent separate stratigraphic units has been the source of considerable disagreement in time past.[1] Nearly everyone, however, who has studied the fossils obtained from these beds has agreed that there are two fairly distinct, though closely related, faunas, one the fauna of the Turritella inezana zone, the other that of the Turritella ocoyana zone.

Recent stratigraphic and paleontological work, the results of which are still unpublished,[2] appears to show that at certain localities in California there were crustal movements of considerable magnitude between the deposition of the Vaqueros and the Temblor. The proper valuation of this hiatus is, in the writer's mind, still an open question. The faunas appear to be fairly closely related, and because of the obscure stratigraphic relations at various localities and the general similarity of the faunas the groups have usually been thrown together. The United States Geological Survey, in its more recent publications on Coast Range geology, applies the name "Monterey group" to these deposits, but the writer considers the Monterey a "series" because, at least in certain localities, it is composed of two epochs of deposition, the Vaqueros and the Temblor.

Stratigraphic relations of the Monterey series to the Upper Oligocene.—There is no conclusive evidence that there were any great

[1] G. D. Louderback, "Monterey Series of California," *Bull. Dept. Geol., Univ. Cal.*, Vol. VII (1913), No. 10, pp. 177–241. F. M. Anderson, "Stratigraphic Study of the Mount Diablo Range of California," *Proc. Cal. Acad. Sci.*, Third Series, Vol. II (1905), No. 2, pp. 161–248. F. M. Anderson, "Further Study of the Mount Diablo Range of California," *Proc. Cal. Acad. Sci.* Fourth Series, Vol. III.

[2] Mapping by Dr. Kew of the United States Geological Survey shows an important unconformity in southern California between the Temblor and the Vaqueros. Mr. Wayne Loel, formerly of Leland Stanford University, is working on a monograph of the Vaqueros. He believes that the faunas of the Temblor and the Vaqueros represent two distinct horizons.

crustal movements just previous to the deposition of the Vaqueros. However, that there was an important hiatus following the deposition of the San Lorenzo is brought out by a comparison of the San Lorenzo and Vaqueros faunas. Very few of the species of the San Lorenzo (Upper Oligocene) have been found in the Vaqueros, while a very large percentage of the species of the latter horizon are common to the Temblor. It is this great faunal change between the San Lorenzo and Vaqueros that is most significant and indicative of one of the major breaks.[1]

Correlation of the Temblor and Vaqueros.—As in the case of the Oligocene, very little direct evidence has been obtained for the correlation, on the basis of the invertebrates, of the divisions of the Monterey series with the Lower-Middle Miocene of the eastern province and Europe. These deposits of the Monterey series were first referred to the Miocene by Conrad[2] as early as 1837. This determination was made chiefly on the general similarity of the generic assemblages to the faunas of the Atlantic Coast Miocene. Following Conrad, the beds here referred to the Monterey series were determined by Whitney and Gabb, both of the old California State Geological Survey, as Miocene. No attempt was made by these pioneers to recognize any subdivisions in the Miocene. Beds now recognized as Upper Miocene (San Pablo) were called Pliocene by Whitney and Gabb[3].

The first announcement of a correlation which gave a fairly definite position to the Temblor group appeared in a paper by Professor J. C. Merriam, entitled "Tertiary Vertebrate Faunas of the North Coalinga Region of California."[4] Previously the Temblor had been referred by some geologists to the Lower Miocene and by others to the Oligocene. In the region of North Coalinga

[1] B. L. Clark, "San Lorenzo Series of Middle California," *Bull. Dept. Geol., Univ. Cal.*, Vol. XI (1918), No. 2, p. 105.

[2] T. A. Conrad, "Fossils from Northwestern America," *Geol. U.S. Ex. Exped.*, Vol. I (1849), App., pp. 723–29, Pls. 17–20; *Proc. Acad. Nat. Sci. Phila.*, Vol. VII (1837), p. 441.

[3] W. M. Gabb, *California Geological Survey: Palaeontology*, Vols. I and II (1864–69).

[4] J. C. Merriam, "Tertiary Vertebrate Faunas of the North Coalinga Region of California," *Trans. Am. Phil. Soc.*, New Series, Vol. XXII (1915), Part III, pp. 1–44.

the land formation locally known as the Big Blue was intercalated with beds of Temblor age from which good marine faunas have been obtained.[1] The Big Blue was first considered by Arnold and Anderson[2] as a part of the Santa Margarita (Upper Miocene). Later mapping, however, showed that it is more closely connected to the Temblor (the so-called Vaqueros) than to the Santa Margarita of that section.

In describing the fauna obtained from the Big Blue, Merriam says:

> In terms of the vertebrate series of Western North America the fauna of the Merychippus zone in the north Coalinga region is clearly later than lower Miocene and not later than upper Miocene. The fact that the Big Blue comes in a section where the Temblor deposits are very thin, and we think are only the top of that section, makes it seem reasonable to believe that the Temblor deposits as a whole belong to the middle Miocene rather than to a part of the upper Miocene.[3]

A clue to the age of the Vaqueros deposits was obtained very recently by the discovery of land-laid deposits near the south end of the San Joaquin Valley which are intercalated with marine deposits of the Vaqueros age. Beds containing a Vaqueros fauna are found immediately below these land-laid beds, and the marine beds immediately above are believed to represent the same horizon. The announcement of the discovery of a vertebrate fauna obtained from these land-laid beds associated with the Vaqueros was recently made by Dr. Chester Stock.[4] These land-laid deposits, referred to as the Tecuja beds, were tentatively correlated by Dr. Stock with the John Day horizon of Oregon. He reports the presence of the genus *Hypertragulus*, a form related to the early camels and deer. *Hypertragulus* occurs both in the Upper Oligocene and the Lower Miocene, but the species from the Tecuja beds seems more

[1] The beds mapped as Vaqueros in the Coalinga field by the United States Geological Survey belong to the Temblor horizon rather than to the Vaqueros.

[2] R. Arnold and R. Anderson, "Geology and Oil Resources of the Coalinga District of California," *U.S. Geol. Surv., Bull. 396* (1909), p. 90.

[3] J. C. Merriam, "Tertiary Vertebrate Faunas of the North Coalinga Region of California," *Trans. Am. Phil. Soc.*, New Series, Vol. XXII (1915), Part III, p. 20.

[4] Chester Stock, "An Early Tertiary Vertebrate Fauna from the Southern Coast Ranges of California," *Bull. Dept. Geol., Univ. Cal.*, Vol. XII (1920), No. 4, pp. 267–76.

closely related to the John Day (Upper Oligocene) form than to that found in the lower Rosebud (Lower Miocene) of the Great Basin region. While Dr. Stock has indicated the relationships of of the Tecuja vertebrate fauna to that of the John Day, he has also stated that it may occupy a position in the Tertiary transitional between Oligocene and Miocene. It seems to the writer that the stratigraphic and paleontologic evidence favors the Lower Miocene age of these beds rather than the Upper Oligocene. The most important evidence in favor of this last conclusion is that the invertebrate fauna of the Vaqueros as already stated, is very closely related to that of the Temblor, a very large percentage of the species being common to the two faunas. Some of the forms listed are types of considerable ornamentation and complexity and are known to have a fairly short geological range. On the other hand, the known fauna of the Vaqueros is very different from the known fauna of the San Lorenzo, and there is here a much greater faunal break than between the Vaqueros and the Temblor. In the section at the south end of the San Joaquin Valley, where the Tecuja beds occur, the Vaqueros rests directly and unconformably upon the San Lorenzo.

Climate.—The conditions of temperature during the Vaqueros and Temblor epoch seem to have been between warm-temperate and subtropical. The generic assemblages of the two horizons are very similar. The large lyropectens and dosinias are found in both, and a fairly high percentage of the species are common to the two horizons. This close relationship of the faunas indicates a similar temperature of the waters.

One of the puzzling problems in connection with the origin of the Temblor deposits is the great thickness of organic shales that is found all along the coast of California and especially in the southern part of the state. In some localities these organic shales, a very large proportion of which are composed of the frustules of marine diatoms, have a thickness exceeding 5,000 feet. Diatomaceous oozes in any considerable quantity are now only found in Arctic and Antarctic waters, and from this we might judge that these shales were deposited in cold water. However, the fossil molluscan faunas found in these shales, or closely associated with them,

indicate a moderate temperature. Dr. J. C. Branner, in a paper read before the Cordilleran Section of the Geological Society of America, suggested an explanation of this apparent disagreement between the floral and faunal evidence.[1] The great thickness of diatomaceous shales in southern California is to be explained by the hypothesis that cold currents carried the diatoms southward along the coast and finally into the partially land-locked basins of southern California where they were killed by the change in temperature. The continuous supply from the north resulted in great thicknesses of deposits composed largely of the tests of these minute plants. This hypothesis may also be an explanation of the origin of the diatomaceous shales of the Oligocene and Upper Miocene.

Paleogeography.—The Temblor deposits have much wider distribution than those of the Vaqueros (Figs. 6 and 7) and are found on the eastern as well as the western side of the Coast ranges. On the eastern side from the vicinity of Coalinga northward these deposits are composed of coarse clastics, while to the west organic shales cover the larger part of the section. The comparison of the area covered by this sea with that which existed during Lower Oligocene time (when the Kreyenhagen shales were deposited), (Fig. 4) shows a marked change. As has already been stated, the Oligocene sediments were deposited in an inland north–south trough very similar to that which existed during the Eocene. The deepest part of the Oligocene trough was on the eastern side of the present Coast ranges, a very large part of the western side at that time apparently having been subject to erosion. On the other hand, the deepest part of the Temblor sea was on the western side of the present Coast ranges; the areas which had been land during the Oligocene were inundated, while to the east, where the Oligocene trough had been deepest, the strand-line deposits indicate shallow water conditions. At this time the interior Diablo range probably formed an archipelago of islands.

San Pablo series.—The San Pablo series is recognized as the second major division of the Miocene. Like the San Lorenzo and

[1] J. C. Branner, "Influence of Wind on the Accumulation of Oil-bearing Rocks," *Proc. Thirteenth Ann. Meeting of the Cordilleran Section of the Geol. Soc. of Am., Bull. Geol. Soc. Am.*, Vol. XXIV (1913), pp. 94–95.

Monterey series, the San Pablo is divisible into minor units on the basis of stratigraphy and fauna. The faunal changes and disconformable relationships of the beds indicate that the sea advanced and retreated three times in middle California, during this period. It is only in middle California that we find the complete sequence of the Upper Miocene series. The two lower divisions of the San

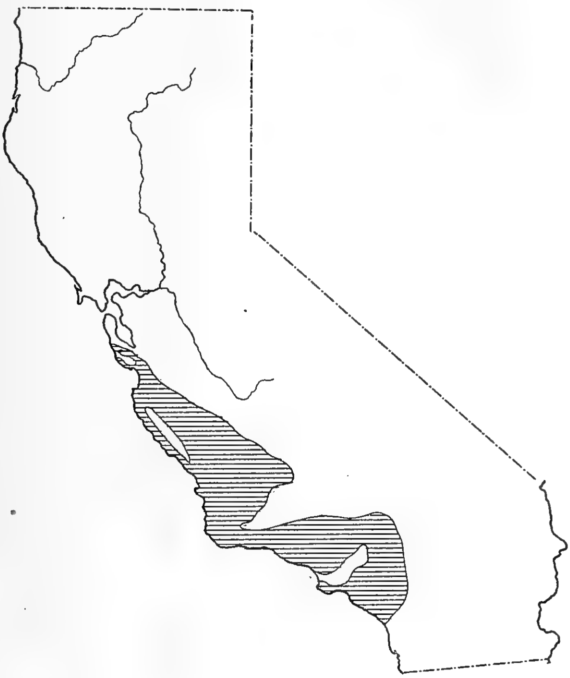

FIG. 6.—Vaqueros (lower Monterey, Miocene)

Pablo series, the Briones and Cierbo groups,[1] have been recognized only in the general region of San Francisco Bay.

Stratigraphic relationships.—In certain sections immediately east of or in the Salinas Valley, a distance of not more than 100 miles

[1] The use of the term San Pablo for the Upper Miocene series of deposits on the West Coast makes it necessary to dispense with the term San Pablo within the group. The name Cierbo is therefore used in this paper in referring to the middle group of the San Pablo series. The type section of the Cierbo is in the south side of the Canada del Cierbo near Carquinez straits. Santa Margarita is a name in common use for the upper portion of the section in the southern part of the state of California and will be applied as a general name for the upper member of the San Pablo series.

from the San Francisco Bay area, we find evidences of crustal movements between the Temblor and the Santa Margarita which have been described as mountain-making. In the Salinas Valley region it is not uncommon to find the difference in dip between the Temblor and the Santa Margarita as much as 30° to 90°, together

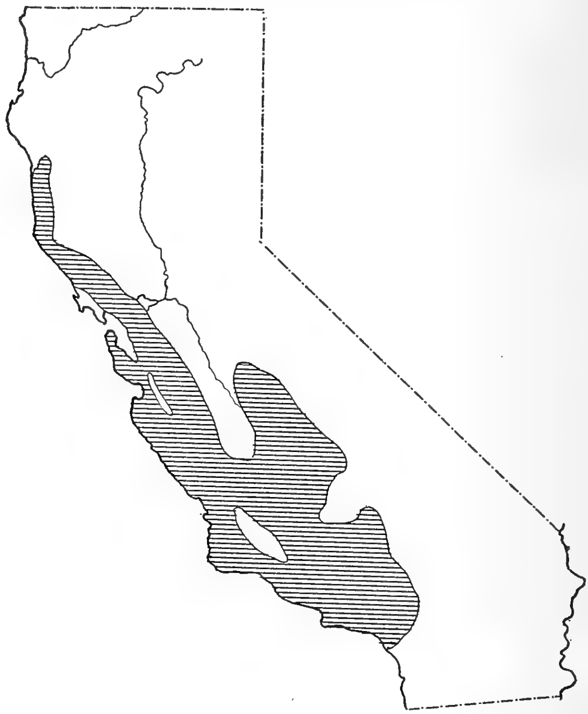

FIG. 7.—Temblor (upper Monterey, Miocene)

with marked difference in strike. Dr. Arnold, in describing the movements that caused this unconformity, says:

One of the most widespread and important periods of diastrophism in the Tertiary history of the Pacific Coast was that immediately following the deposition of the Monterey or lower middle Miocene. Its effects are visible from Puget Sound to southern California It is marked as much by readjustment, by local faulting and folding as by general movements of elevation and subsidence. In some regions the folding and faulting were intense, the greatest disturbances accompanying the uplift of the mountain ranges to an altitude of thousands of feet. In other regions low broad folds were formed during the

post-Monterey disturbance, and the strata were not upheaved to a great altitude. Faulting on a most magnificent scale took place along the earthquake rift and certain other fault-zones, especially that in the Salinas Valley, and along these lines of displacement, masses of granitic rocks, which during the preceding epoch had been subject to little or no erosion, were suddenly thrust upward and left exposed to the ravages of streams that assumed the proportions of torrents in certain regions, as for instance adjacent to the Carrizo Plain in south-central California. The post-Monterey diastrophic movements in the Puget Sound province also produced sharp relief as is evidenced by the coarse sediments immediately following the disturbance. The localization of movements during the period is exemplified at numerous localities in the Coast Ranges.

Throughout much of the coastal belt, and probably likewise in the interior, great volcanic activity took place during the middle Miocene, this being the last epoch of volcanism in the Coast Ranges, south of San Francisco Bay.[1]

It is interesting to note that only a comparatively short distance to the east of the southern Salinas Valley area, where the great unconformity between the Temblor and the Santa Margarita deposits is best seen, it has been very difficult to find the line separating these two horizons. Both the Santa Margarita and Temblor deposits to the southwest of the San Joaquin Valley are composed of organic shales, and in consequence of the difficulty in separating the two horizons the United States Geological Survey has applied the name Maricopa shale to the deposits as a whole.[2] No marked stratigraphic break has been found in middle California between the deposits of the lower San Pablo series (Briones group) and those representing the Temblor. Here the beds of these two horizons are parallel, the chief basis for making the separation being irregular contacts and the difference between the faunas. In middle California, therefore, we have no evidence of crustal movements immediately after the deposition of the Temblor.

Correlation.—Direct evidence for the correlation of the San Pablo series with the eastern and European sections is lacking. The writer has presented the evidence for the correlation of this

[1] R. Arnold, "The Environment of the Tertiary Faunas of the Pacific Coast of the United States," in Willis *et al.*, *Outlines of Geology* (1910), p. 241.

[2] R. W. Pack, "Geology and Oil Resources: Sunset-Midway Oil Field, California," *U.S. Geol. Surv., Prof. Paper 116* (1920), p. 35.

group in a former paper. The correlation is based upon an analysis of the molluscan fauna by the percentage method and the evidence afforded by the occurrence of vertebrates in beds immediately above and below the San Pablo. The following quotations are taken from the above-mentioned paper:

> The percentage of Recent molluscan species in the San Pablo of middle California as listed by the writer is 23 plus; as based upon the gastropods the percentage is only 11 per cent. If we use the percentages as applied to the east coast Neocene and if we can rely upon the equal refinement in the determination of the species, the San Pablo may be considered to be upper Miocene in age, possibly lower Pliocene.
>
> Probably the best evidence showing the age of the uppermost beds of the San Pablo of middle California comes from vertebrate material obtained in the fresh-water beds which in middle California overlie unconformably the San Pablo group. This material was described by Professor J. C. Merriam in his paper "Vertebrate Fauna of the Orinda and the Siesta Beds in middle California."[1] His conclusions as to the age of these beds as shown by the vertebrates are as follows: "The mammalian remains known from both the Orindan and Siestan up to the present time all represent forms such as might be expected in the late Miocene or in the earliest Pliocene, but it will be necessary both to have better material from the Orindan and Siestan and to have well known faunas of western Miocene and Pliocene for comparison before the last word on the age determination can be pronounced.
>
> "Considering the indefiniteness of all the factors concerned, one would not seem justified in being more definite than to state that the Orindan and Siestan faunas are near a late Miocene stage. When the faunas of the two formations are better known, it may appear that more than one stage is represented."[2]

The reader will remember from the discussion of the age of the Temblor that the vertebrate fauna obtained from the Big Blue formation, which is apparently intercalated with the Temblor deposits in the north Coalinga region, was determined by Dr. J. C. Merriam as being not earlier than Middle Miocene. In this same section the Santa Margarita formation is found unconformably above the Big Blue and marine Temblor beds. Also, as will be brought out in the discussion of the Pliocene, a Lower Pliocene vertebrate fauna was found in land-laid beds which rest unconform-

[1] J. C. Merriam, "Vertebrate Fauna of the Orindan and Siestan Beds in Middle California," *Bull. Dept. Geol., Univ. Cal.*, Vol. VII (1913), No. 19, pp. 373–85.

[2] B. L. Clark, "The Fauna of the San Pablo Group of Middle California," *Bull. Dept. Geol., Univ. Cal.*, Vol. VIII (1915), No. 22, p. 439–42.

ably upon the top of the Santa Margarita. Thus it would seem that if we can trust the correlation on the basis of the vertebrates, there can be very little doubt as to the age of the Santa Margarita group which comes between two vertebrate horizons, one not earlier than Middle Miocene, the other not later than Lower Pliocene.

Climate.—The paleontological evidence seems to show that, beginning with the Upper Miocene, there was a temperature differentiation on the West Coast that was even more marked than that existing today.

The Briones and Cierbo groups (lower and middle San Pablo) are not found in southern California, and because of their limited distribution give us very little evidence of temperature differentiation.

The fauna obtained from the Santa Margarita (upper San Pablo) in middle California may be regarded as approximately warm-temperate, and if it were now living it would probably not be found south of Santa Barbara County. This conclusion is based upon the large percentage of recent species found in the faunal assemblage and common to the fauna now found living between San Francisco Bay and Santa Barbara. The presence of certain recent species and the absence of certain genera found at northern localities indicate that the Santa Margarita horizon in southern California represents a warmer facies than that found in middle California. There is, however, a sufficient number of distinctive species common to the two horizons to establish their correlation, though the faunas are on the whole very different.

The fauna of the Montesano formation of Washington, described by Dr. C. E. Weaver,[1] is apparently Upper Miocene in age, but just what part of the San Pablo series it represents has not been established. This fauna, judging from the recent genera and species in the assemblage, is boreal and consequently very different from that of the San Pablo. If the correlation of the Montesano formation with the San Pablo series is correct, there was at that time a temperature differentiation comparable to that found between the recent faunas of middle California and Alaska.

[1] C. E. Weaver, "Tertiary Formations of Western Washington," *Wash. Geol. Surv., Bull. 13* (1916), pp. 1–327.

Paleogeography.—The sediments of the Briones and Cierbo groups were deposited in a limited arm of the sea confined to the San Francisco Bay region (Figs. 8 and 9). Where all the groups of the San Pablo series are present, erosion contacts are found separating them.

With the opening of the Santa Margarita there was a great inundation, somewhat comparable to that of the Temblor, though

FIG. 8.—Briones (lower San Pablo, Miocene)

the basins of the former were more local. Beds of Santa Margarita age are found from a little north of San Francisco to the region just north of Los Angeles (Fig. 10). The deposits are found on both the eastern and the western sides of the Coast ranges, and over large areas to the south of San Francisco these deposits are composed very largely of organic shales of considerable thickness.

PLIOCENE

The marine Pliocene of the West Coast as now recognized is divisible into at least three distinct horizons: the Jacalitos, the Merced, and the Saugus. Continental deposits are found in different parts of the Coast ranges representing all three of these horizons, and in some instances to which reference has already been made these continental deposits are found closely associated with the marine beds.

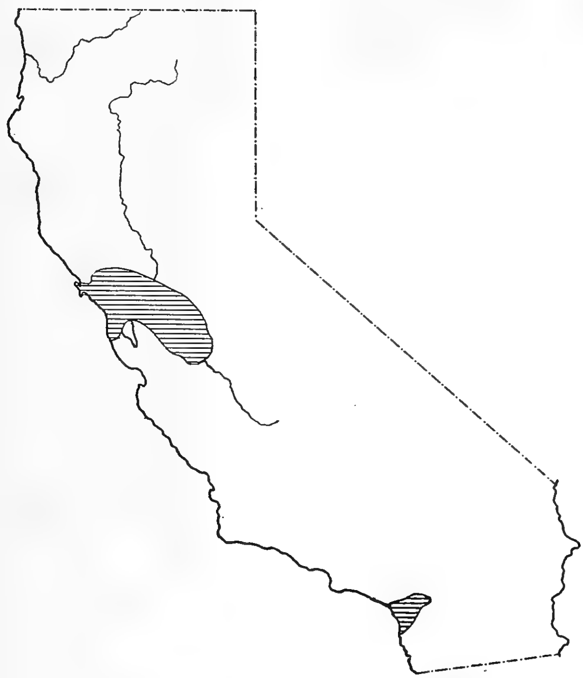

FIG. 9.—Cierbo (middle San Pablo, Miocene)

Stratigraphic relationship of the Pliocene and Upper Miocene.—No evidence of folding between the beds of the Santa Margarita (Upper Miocene) and the Jacalitos (Lower Pliocene) has been obtained in southern California, though good erosion contacts are found separating them and the faunas are on the whole very different. In the vicinity of Mount Diablo, just to the east of the San Francisco region, the Santa Margarita beds were folded before the

Lower Pliocene deposits were laid down. The Lower Pliocene in this region is composed of the Pinole tuff and the Orinda formation which are of continental origin.

Faunal relationships of the Pliocene.—Dr. Nomland's study of the faunas of the Jacalitos ("lower Etchegoin") and Santa Margarita has shown that they are very distinct, and that the hiatus between them was more than local.[1] None of the highly ornamented gastropods, pelecypods, or echinoids has been found common to the two, and the percentage of recent species in the Santa Margarita is much less than that in the Jacalitos (lower Etchegoin) of Nomland.

An unconformity has been found in the Fernando series in the region just north of Los Angeles which is probably the largest and most important stratigraphic break in the West Coast marine Pliocene.[2] Over a fairly large area in that region there is a marked difference in dip and strike between the lower and middle Fernando, now referred to the Pico formation by the United States Geological Survey, and what has previously been referred to as the upper Fernando.[3] The beds of this upper horizon contain a very large percentage of recent species. The Geological Survey proposes to use the name "Saugus formation" for the upper Fernando section. It is herein referred to as the Saugus group. The faunal break between the Saugus and the Pico indicates a great lapse of time. Indeed, the difference is so great that the question may be raised as to whether the Saugus does not belong to the Pleistocene

[1] J. O. Nomland, "Fauna of the Lower Pliocene at Jacalitos Creek and Waltham Canyon, Fresno County, California," *Bull. Dept. Geol., Univ. Cal.*, Vol. IX (1916), No. 14, pp. 199–214; "Fauna of the Santa Margarita Beds in the North Coalinga Region of California," *Bull. Dept. Geol., Univ. Cal.*, Vol. X (1917), No. 18, pp. 293–326.

[2] G. H. Eldridge, and R. Arnold, "Santa Clara Valley, Puente Hills and Los Angeles Oil Districts of California," *U.S. Geol. Surv., Bull. 309* (1907), pp. 1–259. R. Arnold and R. Anderson, "Geology and Oil Resources of the Santa Maria District, California," *U.S. Geol. Surv., Bull. 322* (1907), pp. 1–157. W. S. W. Kew, "Structure and Oil Resources of the Simi Valley, Southern California," *U.S. Geol. Surv., Bull. 691 M* (1919), pp. 323–55.

[3] A paper by Dr. W. S. W. Kew of the United States Geological Survey is now in press in which the Fernando is considered a group composed of the Pico and Saugus formations separated by an unconformity.

rather than to the Pliocene. If so, the West Coast Pleistocene formations have been generally folded.

Correlation.—There has been considerable confusion in times past as to the proper sequence of the Pliocene. The difficulties appear to have been due to two factors: first, the basins of deposition during the Pliocene were more local and isolated than they had

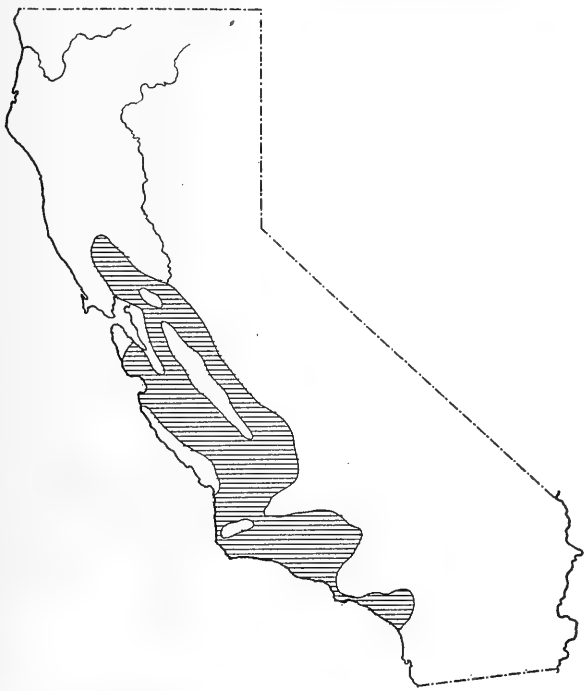

FIG. 10.—Santa Margarita (upper San Pablo, Miocene)

been during previous periods; and second, conditions of temperature varied from one locality to another. These factors were undoubtedly the cause of the great differences found between the faunas of the various provinces; however, the latter factor, temperature, was the most important. The great confusion as to the proper sequence of the Pliocene formations is reflected in the numerous names which have been given to them. Beginning

in Oregon, we have the "Empire" formation;[1] in northern California, the "Wildcat";[2] in the region of San Francisco Bay, the "Merced";[3] and a little to the south of this, in Santa Cruz County, is the "Purissima."[4] Still farther to the south are the "Jacalitos" and "Etchegoin" formations in the Coalinga region;[5] the "Pico" and "Saugus" formations of the Fernando group[6] in the Ventura region and the "San Pedro"[7] and "San Diego" Pliocene deposits on the southern California coast. The "Purissima," "Jacalitos," "Etchegoin," and "lower Fernando" have in the past been referred to the Upper Miocene and correlated with the San Pablo of middle California.

During the last few years our ideas of the sequence of these various formations have been very radically revised as the result of more detailed studies of vertebrate and invertebrate faunas. The writer's study of the San Pablo series convinced him that the faunas of that series belong to an older horizon than those of the Jacalitos-Etchegoin, Purissima, and Pico, and that the percentage of recent species in the latter beds indicate Pliocene age. Later work by Dr. Nomland on the Jacalitos and Etchegoin of the Coalinga region corroborated these conclusions.

[1] W. H. Dall, "The Miocene of Astoria and Coos Bay, Oregon," *U.S. Geol. Surv., Prof. Paper 59* (1909), pp. 1–261.

[2] A. C. Lawson, "The Geomorphogeny of the Coast of Northern California," *Bull. Dept. Geol., Univ. Cal.*, Vol. I, No. 8 (1894), pp. 24–272. Bruce Martin, "Pliocene of Middle and Northern California," *Bull. Dept. Geol., Univ. Cal.*, Vol. IX (1916), No. 15, pp. 215–59.

[3] A. C. Lawson, *U.S. Geol. Surv., Folio 193*, San Francisco Folio; "Post-Pliocene Diastrophism of the Coast of Southern California," *Bull. Dept. Geol., Univ. Cal.*, Vol. I, No. 4 (1893), pp. 115–60.

[4] J. C. Branner, F. G. Newsom, and R. Arnold, *U.S. Geol. Surv., Folio 163*, Santa Cruz Folio.

[5] J. O. Nomland, "Fauna of the Lower Pliocene at Jacalitos Creek and Waltham Canyon, Fresno County, California," *Bull. Dept. Geol., Univ. Cal.*, Vol. IX (1916) No. 14, pp. 199–214; "Etchegoin Pliocene of Middle California," *Bull. Dept. Geol., Univ. Cal.*, Vol. X (1917), No. 14, pp. 191–254. R. Arnold, "Palaeontology of the Coalinga District, California," *U.S. Geol. Surv., Bull. 396* (1909), pp. 5–169.

[6] W. A. English, "Fernando Group near Newhall California," *Bull. Dept. Geol., Univ. Cal.*, Vol. VIII (1914), No. 8, pp. 203–8.

[7] R. Arnold, "Palaeontology and Stratigraphy of the Marine Pliocene and Pleistocene of San Pedro, California," *Cal. Acad. Sci., Memoir III* (1903).

The most conclusive evidence of the Pliocene age of the Jacalitos and Etchegoin was the discovery of fossil land vertebrates in these formations in the Coalinga district. These vertebrate remains were obtained from three distinct horizons: one in what has been mapped as Jacalitos, another in the middle of the type section of the Etchegoin, and the third at the top of the Etchegoin. Professor J. C. Merriam, to whom this material was referred for study, has determined these faunas to be of Pliocene age. Merriam referred to the lowest fauna, that of the Neohipparion zone, as being not older than Lower Pliocene.[1] The next horizon is fairly well up in the Pliocene, and the fauna from the upper Etchegoin is referred to the Upper Pliocene.

It is interesting to note that stratigraphically above the Pliohippus proversus beds of Merriam (uppermost Pliocene), in the above-mentioned section, there are several thousand feet of land-laid deposits which are folded and are older than the Pleistocene terraces of that region. No evidence of the exact age of these beds has been obtained, but it is suggested that they may be the land-laid equivalent of the Saugus of that vicinity.

Climate.—One of the most interesting results of the study of the invertebrate faunas of the Pliocene is the evidence they give of temperature differentiation. The fauna from the Wildcat and Merced of northern and middle California is essentially boreal in character. In southern California the Pliocene is for the most part represented by a fairly warm-temperate fauna. These two faunas, the boreal and the warm-temperate, have very little in common, and consequently it was a long time before their contemporaneity was recognized. The solution of the problem was obtained from the fauna of an intermediate area. The fauna of the type section of the Purissima in the Santa Cruz Mountains of California is in part warm-temperate and in part boreal, and certain species very common in the north, some of which are fairly highly ornamented forms, were found in this section.

Paleogeography.—It was noted in the first paragraph, under the discussion of the factors causing the differentiation of faunas of

[1] J. C. Merriam, "Tertiary Vertebrate Faunas of the North Coalinga Region of California," *Amer. Phil. Soc. Trans.*, N.S. Vol. XII, Pl. III (1915), pp. 26–43.

the Pliocene, that the basins of deposition were local. Crustal movements appear to have been more frequent during the Pliocene than during the other periods of the Tertiary, and we may therefore expect to find a large number of stratigraphic breaks which do not represent any very great time break. At the opening of the Pliocene the Jacalitos sea was much more limited than the seas of

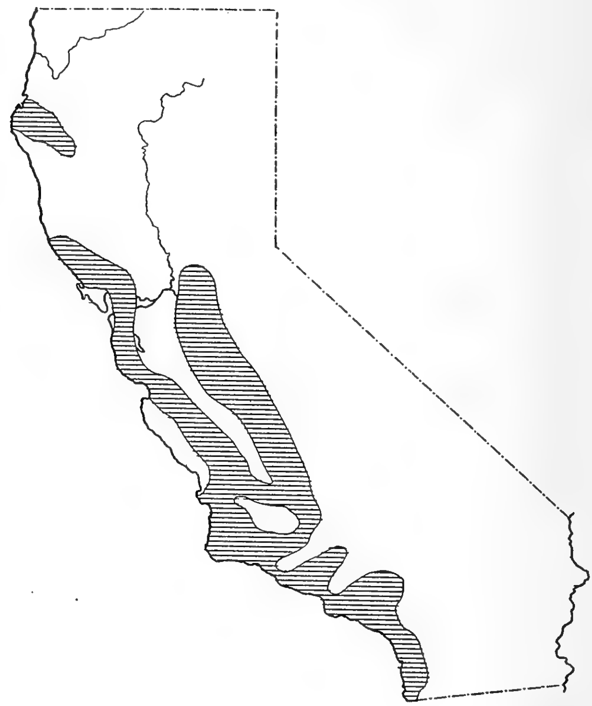

FIG. 11.—Merced (Middle Pliocene)

the Middle or possibly the Upper Pliocene, and the principal localities where deposition took place were probably in the region of the southern and western side of the San Joaquin Valley and the southern part of the Salinas Valley. This sea was probably connected with the ocean in the vicinity of the upper Salinas Valley.

With the opening of the middle Pliocene a great change had taken place. The sea transgressed over wide areas to the north. Areas which had previously been subject to erosion were covered.

This great incursion is referred to here as the Merced sea and represents the time of deposition of the Purissima, Etchegoin, and Jacalitos (Fig. 11). The condition was similar to that which existed in the Upper Miocene, when there were restricted basins at the opening of the period, a great incursion during the later part of the period, and then a retreat shown by widespread unconformities.

FIG. 12.—Saugus (Upper Pliocene)

The Saugus formation, of which little is known, appears to have been formed after the folding and consequent retreat of the sea from the great inland basins in the San Joaquin and Salinas Valley districts. The Saugus beds are largely confined to the coast, but in the Ventura district an embayment extended a considerable distance inland (Fig. 12).

The Pliocene period was followed by great crustal movements which folded the formations in great anticlines and synclines, some

of which were overturned. In certain localities the Cretaceous has been thrust over the younger formations. Though this great period of crustal movement had its culmination at the close of Saugus time, it was well under way even prior to the deposition of these beds. In the past geologists have generally assumed that the division between the Pliocene and Pleistocene occurs after this great folding, that is, at the end of Saugus time. But the very large percentage of the Recent species in the Saugus suggests that the group may possibly be of Pleistocene age, and therefore a large part of the folding which created the Coast ranges may have taken place during Pleistocene time.

OUTLINE OF PLEISTOCENE HISTORY OF MISSISSIPPI VALLEY[1]

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RELATION OF PRESENT STREAM TO PREGLACIAL VALLEYS

The headwater portion of the Mississippi River, above St. Paul, Minnesota, is in a region so thickly covered by glacial deposits that the present streams are entirely independent of the preglacial valleys, and their history begins with the recession of the ice in the last, or Wisconsin, stage of glaciation. The courses of preglacial drainage lines in this region have been only partially traced by means of deep borings. This paper deals, therefore, mainly with the part below St. Paul.

For 15 miles below St. Paul the Mississippi follows the valley of a small tributary of the preglacial river, the course of the main valley being a few miles to the west, passing beneath Lake Minnetonka, and across the lower end of Minnesota Valley, and continuing through Dakota County to the present stream at Pine Bend, 6 miles above Hastings. From Hastings to Clinton, Iowa, the river practically follows the course of the preglacial valley, though it cuts off projecting points from the west bluff at Trempealeau, Wisconsin, and in the north part of Clinton, Iowa (Fig. 1).

At the mouth of the Wapsipinicon River, below Clinton, the Mississippi leaves the preglacial valley, passes into a rock gorge, and flows across rapids to Rock Island. Two courses for the preglacial valley have been suggested, one to the southwest from the lower part of Wapsipinicon Valley to Muscatine, on the present Mississippi, the other to the southeast to Hennepin, on the Illinois Valley. Along both lines the glacial deposits are very thick, and the rock surface much lower than the present streams. The southeastward course seems on the whole more likely to have been

[1] Published by permission of Director, U.S. Geological Survey.

FIG. 1.—Map of Mississippi River

followed by the preglacial Mississippi, for other preglacial valleys converge toward it, one followed by the lower course of the Rock River, reversed, and another by Duck Creek, and its continuation in an abandoned valley to the east of Rock Island. The breadth of the valley leading to the Illinois also appears to be greater than that of the one to the southwest. A different stream from that which opened the upper Mississippi Valley probably developed the valley now utilized by the Mississippi below Muscatine. It is likely to have taken the drainage of much of eastern Iowa and drained the region now tributary to Iowa River, though it does not follow the course of that stream. It may, perhaps, be appropriately called the preglacial Iowa River.

The two drainage lines came together at the mouth of the Illinois River. The western one departed slightly from the present Mississippi in the southeast corner of Iowa, the old course being southwest from Fort Madison to the mouth of the Des Moines River, while the present stream turns southward across the Des Moines rapids to Keokuk, and joins the old valley at Warsaw, Illinois.

The present river follows essentially the course of the preglacial valley from the mouth of the Illinois River to Thebes in southern Illinois, though it has cut off points from the west bluff near Grand Tower, which in high-water stages stand as islands in the valley. The Mississippi turns into a rock gorge at Thebes, and a few miles below comes into the old Ohio Valley, while the old valley runs southwestward, and unites with the Ohio Valley farther down.

RELATION OF PRESENT STREAM BED TO BURIED ROCK FLOOR

The present stream falls from 683 feet A.T. at St. Paul to 566 feet at Clinton, or 117 feet in about 300 miles. This section includes Lake Pepin which holds a level of 664 feet for 25 miles. Aside from this lake the stream has an average fall of about 5 inches per mile. The fall of the buried rock floor is very similar, for it stands 480 to 500 feet A.T. near St. Paul and is below 400 feet in the vicinity of Clinton. The rock floor has more or less inequality because of variations in the scour of the stream, and in the hardness

of the bed, so its slope conforms only in a general way with that of the stream which occupied it.

On the line of the old valley leading from Clinton southeastward to the Illinois River at Hennepin the rock floor falls to about 340 feet, as shown by borings at Princeton and Bureau Junction, near Hennepin. The Illinois River at Hennepin is 432 feet A.T. or about 90 feet above the rock floor.

The Illinois River has a fall of only 27 feet in 207 miles from Hennepin to the junction with the Mississippi River, or 1.56 inches per mile. From the mouth of the Illinois River to St. Louis, a distance of 41.5 miles, the Mississippi falls 21 feet, or 6 inches to the mile. The rock floor, as shown by borings and bridge excavations opposite St. Louis, is about 100 feet below the low-water level of the river, or 280 feet A.T. This gives a fall of 60 feet from Hennepin to St. Louis in 248 miles, or about 3 inches per mile.

The distance by the course of the present Mississippi from Clinton to St. Louis is about the same as along the line to Hennepin and down the Illinois, being not far from 300 miles. The fall is about 180 feet, of which 48 feet is over rock rapids in a distance of 24 miles. The remaining 132 feet is at the rate of less than 6 inches per mile. The rock floor in the valley below Muscatine is known to be 100 feet or more below low water. It has been best tested at Fort Madison, and found to be 120 to 135 feet below low water, or 365 to 380 feet A.T. There is thus a fall of nearly 100 feet between Fort Madison and St. Louis in about 200 miles, or a rate very similar to that of the present stream, aside from the rapids.

The fact that these buried valleys have beds 100 to 180 feet below the low-water level of the present streams has led certain geologists to infer that the altitude must have been higher than now when they were excavated. There are, however, certain other conditions that have had influence in causing this difference. One of these conditions is a marked lengthening of the lower course of the Mississippi, or extension of its mouth gulfward, in the long period since the river was flowing on this rock bed. Warren Upham has presented evidence from a study of old maps of the lower course of the Mississippi that the delta has made a marked

extension in the past 400 years.[1] The time since glaciation came in to displace the streams from their preglacial beds is likely to be at least 1,000 times, and perhaps 2,000 times, as long as the 400 years involved in the growth noted by Upham. The extension may, therefore, be as great as from the site of Vicksburg, and perhaps from farther up. With this lengthening there would come a corresponding aggradation of the valley causing the stream to flow at a level materially higher than its rock bed. Studies by the present writer near Osceola in the northeast part of Arkansas have shown that the flood plain has been built up about 35 feet by a sediment finer than the underlying sand presumably in Pleistocene time. So the aggradation there appears to have been at least that amount, and it would be fully as great farther up the valley.

Another condition that affects the Mississippi Valley below the mouth of the Missouri River, results from the greater amount of sediment now brought in by that stream and carried down the Mississippi, as compared with that likely to have been carried by it before its watershed became so extensively covered with loess, for loess furnishes the main part of the sediment. With increase of load a stream raises its gradient. The Missouri has a fall of about 300 feet in its lower 300 miles, while the section of the Mississippi which is less heavily loaded with sediment has a fall of only about 150 feet. Below their junction the stream has a slightly higher rate of fall than that of the Mississippi above the junction, it being 7.2 inches per mile from the mouth of the Missouri to the mouth of the Ohio. The effect of this load of sediment is clearly brought out by comparing this section of the Mississippi with the section of the Ohio immediately above its mouth. The rate of fall in the lower 200 miles of the Ohio is only about 3 inches per mile, or less than half that of the Mississippi; yet the Ohio is a smaller stream, and should have a higher gradient were it carrying a similar load. There is a difference of about 70 feet between the fall of the Mississippi and of the Ohio in the 200 miles above their junction, and this may give a measure of the amount of aggradation at the mouth of the Missouri, due to the heavy load

[1] "Growth of the Mississippi delta," *American Geologist*, Vol. XXX (1902), pp. 103–11.

of sediment. If to this is added the 35 feet of aggradation that is due to the lengthening of the stream, the full difference between the level of the rock bed and the level of the present river is accounted for. There thus seems no need for postulating disatrophic movement. Farther up the river the rock barriers in the path of the stream, above Keokuk and above Rock Island, contribute in holding the present stream to a high level.

GLACIATION IN ITS RELATION TO DRAINAGE MODIFICATION

The preglacial Mississippi in its course from near Red Wing in southeastern Minnesota to the lower Illinois valley, and thence to St. Louis and Cairo, appears to have suffered little or no disturbance in the first two stages of glaciation, the Nebraskan, or pre-Kansan, and the Kansan. The Iowa branch of the preglacial Mississippi was more seriously disturbed by these early glaciations. Its valley was so completely filled as far down as Muscatine that its course is known only by borings. It was also completely filled for a few miles in the southeastern corner of Iowa. The filling as far down as the line of Iowa and Missouri is found to embrace the pre-Kansan as well as Kansan drift. In the vicinity of Fort Madison, Iowa, the pre-Kansan drift fills the old valley from the level of the rock floor, 135 feet below the present stream, to a height of about 75 feet above the river. A black soil marks its upper limit. It is overlain by Kansan drift, and this in turn by Illinoian drift, which rests on a soil developed on the Kansan drift. Other records in eastern Iowa, southern Minnesota and western Wisconsin show that valley excavation had reached its lowest limits prior to the earliest glaciation. One of the clearest records is at Washington, Iowa, brought to notice by Calvin.[1] It is in a locality where no drift later than the Kansan is present, and is in the line of a western tributary of the preglacial valley utilized by the Mississippi below Muscatine. The Kansan drift extends to a depth of 115 feet, and rests on a peaty bed containing wood and cones of the black spruce. Below this is the pre-Kansan drift, extending to a depth of 350 feet from the surface, or to about 400 feet above sea-level. This level of the rock floor fits in well

[1] *American Geologist*, Vol. I (1888), p. 28.

with that of 365 feet at Fort Madison. Attention is directed particularly to this evidence that the lowest valley excavation is preglacial, since Trowbridge, of the University of Iowa, has for some years been advocating that a large part of the valley deepening in the upper Mississippi region was accomplished between the pre-Kansan and Kansan stages of glaciation.[1]

The Kansan drift extends beyond the present Mississippi Valley into the western edge of Illinois, but does not appear to reach the Illinois Valley. Its southern limits on the border of the Mississippi Valley are at the city of St. Louis. The amount of filling with glacial deposits is difficult to determine because of the great amount of erosion since they were laid down. It was sufficient in southeastern Iowa to prevent the stream from occupying the old valley again. But below the mouth of the Des Moines it was reoccupied by drainage when the ice of the Kansan stage melted away. The course of post-Kansan drainage from the part of the valley between Muscatine and the Des Moines rapids is more difficult to determine, for it was later occupied by ice in the Illinoian stage of glaciation. Possibly it drained eastward from Muscatine, and together with the upper Mississippi discharged through the lower Illinois Valley. It is probable that nearly all the erosion of the gorge at the Des Moines rapids has taken place since the Illinoian stage. A comparison of the valley to the east from Muscatine with that across the Des Moines rapids indicates that it was opened earlier. Its rock bed has been cut below the level of the present Mississippi, and it is a broader valley with gentler slopes than that across the rapids. The difference may be seen by comparing charts 148 and 149 of the Mississippi River Commission, embracing the part east of Muscatine, with charts 136 and 137, which embrace the rapids. The valley east from Muscatine is also embraced in the Edgington and Milan quadrangles of the U.S. Geological Survey. The opening of this valley eastward from Muscatine may date from the Nebraskan stage and have been occupied by east-flowing drainage down to the Illinoian stage, when it was changed to a west-flowing stream.

[1] *Proc. Iowa Academy of Science*, Vol. XXI (1915), pp. 208–9; *University of Iowa Studies*, Vol. IX (1921), pp. 123–27.

It seems probable that the Illinoian glaciation was the first to throw the upper Mississippi out of its course through the Illinois Valley. At the culmination of the Illinoian stage of glaciation the Mississippi opened a temporary course across eastern Iowa, just outside the limits of the ice sheet.[1] This connected with the present Mississippi at the head of the gorge across the Des Moines rapids below Fort Madison at a level about 120 feet above the present stream; so nearly all the cutting in rock at these rapids has been done since that time. The Illinoian drift extends a few miles beyond the present course of the Mississippi from Clinton to Fort Madison, but farther north and south its border lies east of the river. With the recession of the Illinoian ice sheet the Mississippi shifted to its present course between Clinton and Fort Madison.

It seems to have been at this time that the course across the rapids above Rock Island was initiated, the ice being still present on the lower land to the east. The channel across these rapids, like that across the Des Moines rapids, has the appearance of being much younger than that between Muscatine and Rock Island.

Goose Lake channel, which branches off from the present Mississippi at the mouth of Maquoketa River and passes southward across Clinton County, Iowa, to the Wapsipinicon Valley, was probably opened in Illinoian time, and may have continued down to Wisconsin time. It seems, however, to have been abandoned before the waters of the glacial lakes began their work of excavation discussed below.

The glacial drainage outside the Illinoian ice sheet in southeastern Iowa carried enough material into the Mississippi Valley below the Des Moines rapids to fill it to a level about 100 feet above the present low-water level of the river. But within a few miles it dropped to a level so nearly the same as that of the filling at the succeeding, Wisconsin, stage of glaciation that it cannot well be differentiated. In each the level of filling is about 60 feet above present low water, and 30 to 40 feet above flood stages.

[1] Frank Leverett, "The Illinois Glacial Lobe," *U.S. Geol. Surv.*, Monograph XXXVIII (1899), pp. 89–97, Pl. VI.

It is probable that the aggradation at the Illinoian stage was sufficient to enable the river to make use of a low pass across the rock ridge below Thebes, Illinois, and thus join the Ohio more directly than by the old course. The contours of this new course seem to be about the same as in the Des Moines rapids and those above Rock Island, and thus to favor a similar age.

The Iowan glaciation appears to have been too light and transitory, or to have encroached on the Mississippi Valley too little, to produce any permanent deflections of the stream. It may, however, have reached to the valley both from the east and west in the vicinity of Clinton, as indicated by maps and descriptions in Monograph XXXVIII.[1] There are deposits of sandy gravel along some of the tributaries of the Mississippi in southeastern Minnesota and in western Wisconsin that are outside the glacial drainage lines which were connected with the Wisconsin ice sheet, and which are tentatively referred to the Iowan glacial drainage. The material is nearly as fresh as gravel of Wisconsin age, and its preservation from erosion is but little different from that of Wisconsin deposits. It seems, therefore, to be too young to refer to the Illinoian glacial drainage.

The Wisconsin glaciation covered only the headwaters of the Mississippi down to a point a little below St. Paul. But it covered the lower Illinois, and built great moraines to the north and west of Hennepin that have effectually barred the upper Mississippi from returning to its old course into the lower Illinois Valley.

The glacial drainage into the Mississippi from the Wisconsin drift at first had its head on the outer slope of a moraine that crosses the river in the southeast part of the St. Paul quadrangle. The drainage was extended up the valley step by step with the recession of the ice border. At the border of the Wisconsin drift there is an extensive outwash plain, 940 to 960 feet above sea level, or 260 to 280 feet above the Mississippi River. From this outwash plain there was drainage down the course of the Vermillion Valley as well as down the present Mississippi, with a descent of fully 100 feet in about 20 miles. The filling is a fine, sandy gravel, and

[1] *Op. cit.*, pp. 131–153, Pls. VI and XII.

well records indicate that it was built up from near the level of the present river. At Red Wing the filling is 125 feet above the river, or 790 feet above sea-level. Its slope is more rapid than the fall of the river as far down as the mouth of the Wisconsin River. It is there about 70 feet above the river, or 675 feet A.T. At Bagley, Wisconsin, 12 miles farther down, the filling is only 60 feet above the river, and it maintains a height of 50 to 60 feet above the river from there down to about the mouth of the Ohio River. Just above the Des Moines rapids, in the vicinity of Fort Madison, the Wisconsin filling is exceptionally well preserved at a level 50 to 55 feet above the river.

There were accessions from tributaries at several points, the most important being St. Croix, Chippewa, Wisconsin, Rock and Illinois rivers. The Illinois is likely to have made a contribution of sediment to the lower Mississippi larger than that brought down from the upper Mississippi. The Illinois Valley was built up at the Wisconsin drift border, at Peoria, 170 feet above the present stream. In 175 miles from Peoria to Alton, at the head of the American Bottoms on the Mississippi, there is a descent of about 160 feet in the surface of the sandy filling of Wisconsin age, while the present stream falls only about 25 feet in the same distance. Below St. Louis there is sandy filling, which seems referable to the Wisconsin glacial drainage, in small remnants in recesses of the valley and in the mouths of tributary valleys. Filling of this sort is found as far down as the place where the Mississippi turns into the gorge near Thebes. The new topographic map of the Jonesboro quadrangle brings out terraces at the mouths of Dutch Creek and Clear Creek valleys which catch the 380-foot contour. They are 30 feet above the broad flood plain of the river.

DRAINAGE FROM THE GLACIAL LAKES

The Mississippi Valley was the line of discharge to the Gulf of Mexico for several of the great glacial lakes that were developed in front of the receding ice border. Lake Agassiz, with a maximum area as large as that of all the present Great Laurentian lakes, drained into the Mississippi through the Minnesota Valley; Lake Duluth, in the Superior Basin, drained through the St. Croix

Valley, and Lake Chicago, in the Michigan Basin, through the Illinois Valley. At times Lake Chicago carried the drainage of glacial lakes in the Huron and Erie basins. As these lakes were settling basins, their outlets carried very little sediment, and the volume of water being large, they cut deeply into deposits that had been

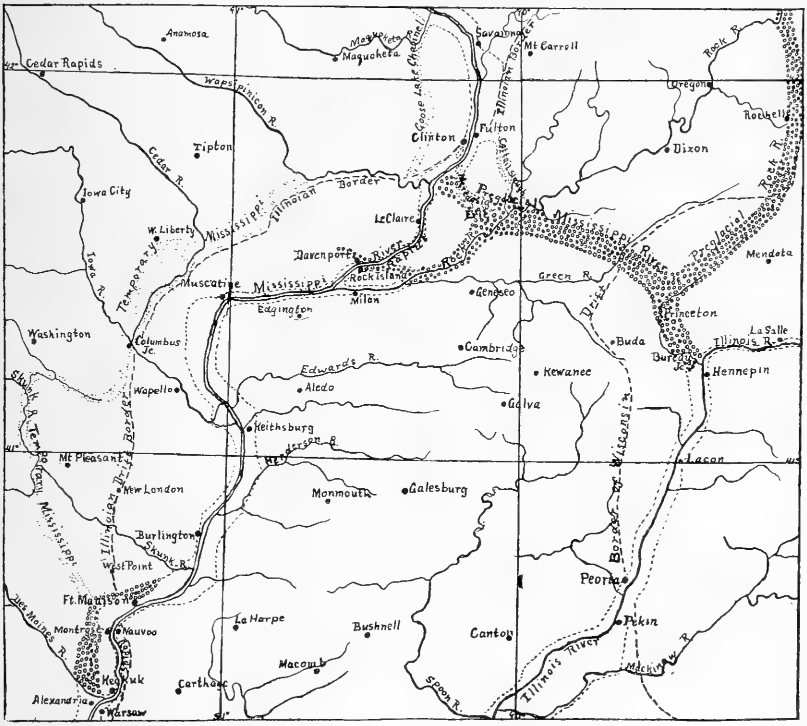

FIG 2.—Map showing shiftings of course of Mississippi River

built up by the waters flowing directly from the ice sheet, and developed a low gradient, lower than is consistent with the present Minnesota, St. Croix, and Illinois rivers, or even with the Mississippi River. At the junction of the Minnesota and the Mississippi it cut to a level 30–40 feet below the present streams, and there is now backwater of that depth in the lower end of the Minnesota Valley, caused by the aggrading effect of the Mississippi River. The conditions are similar at the mouth of the St. Croix

River, and its lower part, below Stillwater, is known as Lake St. Croix. A few miles farther down is Lake Pepin, where for a length of 25 miles the Mississippi is ponded by a barrier built up from sediment brought in by the Chippewa and Zumbro rivers. The level has thus been raised at least 40 feet at the lower end of the lake. There appears to have been aggradation by the present stream along much of the course from Lake Pepin to the head of the rapids below the mouth of the Wapsipinicon.

The channel across these rapids seems to have been inadequate to carry the great volume of water from the glacial lakes, for part of it flowed southward to Rock River through channels now occupied by Meredosia and Cattail sloughs (Fig. 2), and thence down Rock River Valley to the Mississippi below the rapids. Peat deposits now fill these channels to such height that only the highest floods pass through them, but the beds of the channels are about as low as the low-water level of the present river. The rapids thus seem to have been cut down but little since the waters of the glacial lakes ceased flowing over them. The Des Moines rapids, above Keokuk, took the entire flow of the glacial lake waters. The river at the head of the rapids is 50–60 feet lower than the surface of the Wisconsin deposits of sandy gravel immediately above them. It is probable that much of this lowering is due to the waters of the glacial lakes.

The Illinois Valley was cut to a very low gradient by the waters of Lake Chicago. Under present conditions there is a fall of slightly less than 30 feet in the 220 miles below La Salle. But as the mouth of the Illinois is immediately above the mouth of the Missouri the sediment from that stream may have raised the level of the mouth of the Illinois since the waters of Lake Chicago ceased flowing down the Illinois Valley. The amount of aggradation thus produced is a difficult matter to estimate. Possibly it is as great as that below Lake Pepin, some 30 to 40 feet. Attention was directed above to the influence of the sediment of the Missouri on the gradient of the Mississippi below its mouth.

THE PHYSICAL CHEMISTRY OF THE CRYSTALLIZATION AND MAGMATIC DIFFERENTIATION OF IGNEOUS ROCKS

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IV

MAGNETITE

Magnetite and olivine (*with fayalite*).—As discussed in "Silikatschmelzlösungen," I (1903), the individualization boundary between magnetite and fayalite (reckoned by weight) lies at about:

$$0.2-0.25\ Fe_3O_4:0.8-0.75\ Fe_2SiO_4.$$

Near the same boundary, probably with somewhat less magnetite and somewhat more olivine, we find in the "titanomagnetite"-olivinites, chiefly consisting of the so-called "titanomagnetite," which is a mechanical mixture of magnetite and ilmenite, and of olivine, the latter with proportions intermediate between Fe_2SiO_4 and Mg_2SiO_4 (partly stoechiometrically about $0.4\ Fe_2SiO_4 \cdot 0.6\ Mg_2SiO_4$, partly still more Fe_2SiO_4 and less Mg_2SiO_4). As described several years ago by earlier investigators, among them also myself, the olivine in these rocks appears partly porphyric in the "titanomagnetite" when it is quite abundant. As an example we may take the "titanomagnetite"-olivinite from Cumberland in Rhode Island, according to C. H. Warren[1] with an average mineralogical composition of 46.1 per cent olivine ($0.39\ Fe_2SiO_4 \cdot 0.61\ Mg_2SiO_4$), 20.7 magnetite, 18.6 ilmenite, 9.2 labradorite (Ab_3An_4), 3.6 spinel (and in addition a little Or and sulphide).

The specimen at my disposal (Fig. 27) gives the same proportions between olivine and magnetite plus ilmenite, but less labradorite and less spinel. The olivine shows a more or less well-developed idiomorphic contour against the iron-ore minerals.

[1] *Amer. Jour. of Sci.* (1908), p. 175.

A few quite small individuals of magnetite or ilmenite also appear in the olivine.

The sequence of crystallization cannot here be explained in detail, but so much can be said, that the olivine commenced crystallizing at a very early stage, and certainly not much magnetite

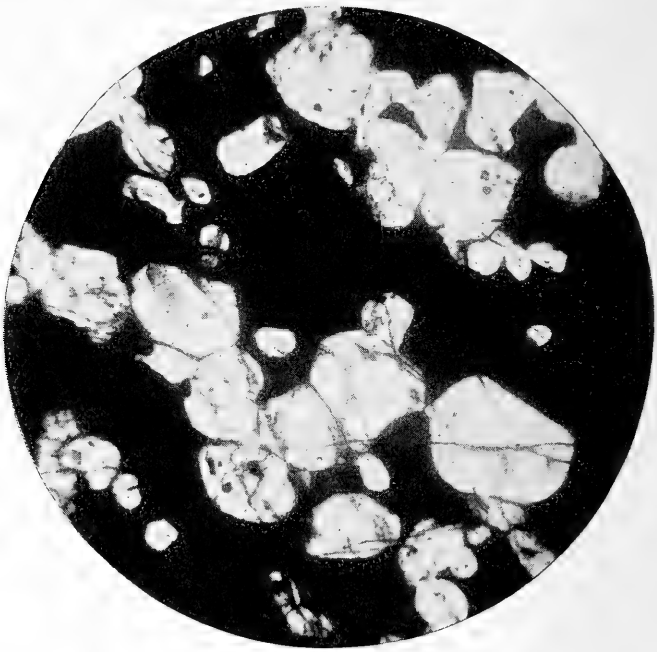

FIG. 27.—Photomicrograph (25:1) of "titanomagnetite"-olivinite from Cumberland, Rhode Island. White = olivine, black = "titanomagnetite."

or ilmenite can have been solidified before the commencement of crystallization of the olivine.

Very interesting in structural respect is a "titanomagnetite"-olivinite (Fig. 28) from Fiskaa in Söndmöre, Norway, consisting of about 50 per cent olivine (according to optical investigations 0.35–0.4 Fe_2SiO_4:0.65–0.6 Mg_2SiO_4); about 40 per cent magnetite plus ilmenite, about 5 per cent spinel; and in addition locally a little hypersthene.

The olivine, with no sign of an idiomorphic contour, shows in nearly every individual an intimate intergrowth with the iron ore,

chiefly magnetite, giving in this manner a form of structure nearly identical with graphic granite or with coarse-grained granophyre. This must be due to a *simultaneous* crystallization of the olivine and the iron ore, especially the magnetite, and the proportion by weight in the intergrown individuals gives about 0.25

FIG. 28.—Photomicrograph (25:1). "Titanomagnetite"-olivinite from Fiskaa, Söndmöre, Norway. White = olivine, black = titanomagnetite. The small gray grains in the photograph are spinel.

magnetite:0.75 olivine, this representing a point or a line on a quite complicated eutectic boundary curve.

Magnetite in gabbro, norite, anorthosite, etc.—As is well known, and still believed by many petrographers, Rosenbusch made the assertion that the oxidic iron ores, magnetite, ilmenite, etc., always belong to the oldest segregations, and crystallized before the solidification of the silicates. This assertion, however, is not correct with regard to the gabbroic rocks, for when only little magnetite (and ilmenite) is present, the crystallization of iron ore does not

commence until the surplus silicate mineral has solidified, so that the quantity of Fe_3O_4 (and $FeTiO_3$) in the rest of the magma has reached a certain amount.

In this connection we choose as an example a rock with a surplus of *plagioclase.* This has previously been described in detail by myself, and treated above, viz., the porphyritic labradorite-norite from Flakstadöen in Lofoten. In proportion to the whole rock 23 per cent labradorite was solidified first, and the crystallization of iron ore, simultaneously with the continued crystallization of plagioclase only commenced when the quantity of iron ore had reached 8.1 per cent Fe_3O_4+0.9 per cent $FeTiO_3$. If we leave out of consideration the components present in small quantity (biotite, apatite, and ilmenite), we obtain the following figures for the commencement of crystallization of the magnetite: about 8.5 per cent magenetite; about 64.5 labradorite; about 13.5 hypersthene; about 13.5 diallage. This represents a point or a line on the individualization boundary between the labradorite and magnetite in a very complicated system, magnetite:labradorite (42 Ab, 6 Or, 52 An):hypersthene:diallage (the two last with many separate components).

In hyperitic-structured gabbro and norite (with olivine gabbro and norite), where at an early stage *much plagioclase* had solidified in the well-known lath-shaped individuals, the magnetite (and ilmenite), when present only in a quantity as small as 1–2 per cent, shows no sign of idiomorphic contour.

On the contrary, the magnetite here only appears as an intervening mass chiefly between the plagioclase individuals (see Figs. 29 and 30), indicating that much plagioclase had already solidified before the magnetite commenced forming.

In this connection we refer to a treatise by S. Foslie (Kristiania) on "The Titanic Iron Ore Deposit at Ramsöy and Its Processes of Differentiation,"[1] where is described a hornblende gabbro, containing about 58 per cent plagioclase, 2.5 per cent magnetite, 31 per cent hornblende, 6.5 per cent biotite, and 2 per cent quartz, with the magnetite formed later than the plagioclase. His photomicrograph, Table I, Figure 1, accords quite well with Figures 29–30 in this paper.

[1] *Geol. Survey of Norway, Aarbog for 1913.*

In the norites, relatively *rich in hypersthene* but quite poor in magnetite (and ilmenite), the crystallization of iron ore did not commence until some part of the hypersthene had solidified. As an example we choose a norite from Skougen in Bamle, which, according to the chemical analysis (see the section on norite in Part II), contains 1.09 per cent TiO_2, 1.44 Fe_2O_3, and 9.42 FeO. The mineralogical composition is about 47 per cent hypersthene (with a little secondary hornblende), 3 per cent biotite, 48 per cent labradorite, and about 1–2 per cent iron ore, and in addition a little

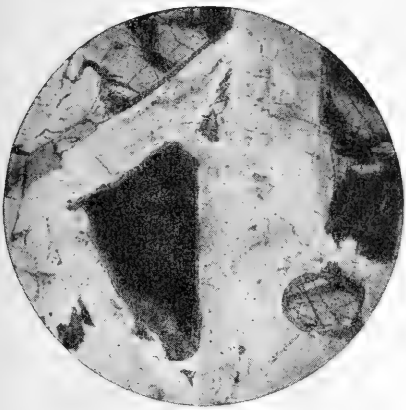

FIG. 29.—Photomicrograph (21:1)

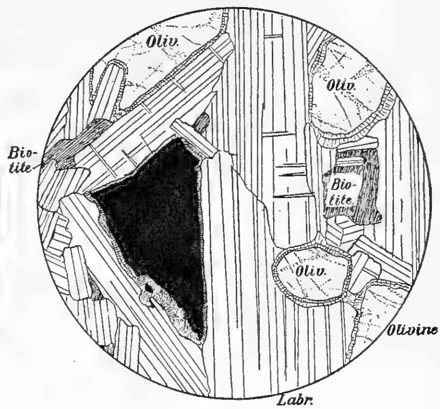

FIG. 30.—Drawing (21:1)

Hyperitic-structured olivine-gabbro from Elverum, Norway. Consisting of *ca.* 65 per cent labradorite, 10 per cent olivine, 10–15 per cent diallage, 3–5 per cent biotite, and *ca.* 2 per cent magnetite (with a little ilmenite).

apatite and pyrite. An essential part of TiO_2 and Fe_2O_3 enters into hypersthene and biotite. In most parts of the thin section (see Fig. 13) iron ore is lacking as it only appears in a few places.[1] Here the hypersthene shows an idiomorphic contour against the magnetite, and this applies especially to the relatively *small* individuals of hypersthene, which partly have an entirely straight crystallographic boundary against the magnetite (with ilmenite).

The latter, on the contrary, shows no signs of idiomorphic outlines. Some part of the hypersthene must thus have solidified before the formation of the magnetite. Since especially the *small*

[1] For this reason we are not able to determine the quantity of the iron ore with precision by a planimeter calculation of the thin section.

individuals of hypersthene show marked idiomorphism against the magnetite (and ilmenite), the crystallization of the iron ore seems to have begun when only the lesser part of the hypersthene had solidified.

Further we find here and there in the same thin section some small crystals of pyrite, surrounded by a thin wreath of magnetite (or "titanomagnetite"). The pyrite accordingly has here served as *Fixkörper* for the deposit of magnetite, which also indicates a solidification of the magnetite at a relatively early stage, while the essential part of the rock still remained in the liquid phase.

Fig. 31.—Photomicrograph (30:1)

Fig. 32.—Photomicrograph (45:1)

From the same thin section of norite from Skougen shown in Figure 13, containing 1–2 per cent "titanomagnetite." Figures 31 and 32 represent two places with much magnetite (and ilmenite), besides hypersthene, a little hornblende, and biotite.

In the olivine-hyperites, *rich in olivine*, with about 25–30 per cent olivine, only 1–2 per cent "titanomagnetite," much plagioclase (labradorite, quite rich in An), and some diallage, etc.; the crystallization, as mentioned above, commenced with the solidification of some olivine, and only then did the "titanomagnetite" commence crystallizing as a deposit on the olivine crystals which acted as a *Fixkörper*. We refer to Figure 33, drawn with the detailed help of a photomicrograph.

We have here treated several gabbros and norites with only *very little* magnetite (and ilmenite). In the corresponding rocks

with a somewhat *larger quantity* of magnetite (and ilmenite) the sequence of crystallization, on the contrary, is different. Here we find idiomorphic crystals, especially octahedrons of magnetite, appearing in the silicate minerals first crystallized. In gabbroidic rocks with much hypersthene, diallage, or olivine, the magnetite seems already to have commenced its crystallization at about 4 per cent Fe_3O_4, and about simultaneously with the first individualized

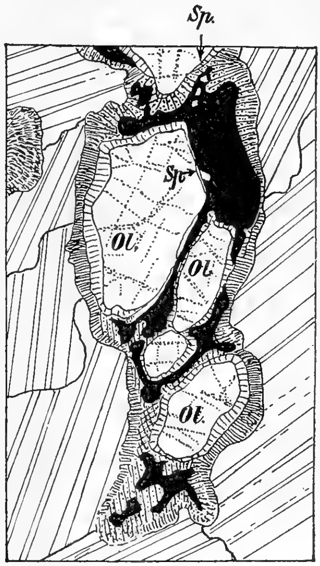

FIG. 33.—Hyperitic-structured olivine-gabbro from Langö, near Kragerö, Norway. Contains about 25 per cent olivine (Ol.), *ca.* 1–2 per cent magnetite with ilmenite (black), 60 per cent labradorite, 12 per cent diallage, 2 per cent brown hornblende, 0.01–0.02 per cent spinel (Sp.) in the magnetite, very little apatite. Drawn partly from photograph and partly from nature.

Magnetite (with ilmenite), younger than the olivine. Olivine with reaction rims against the labradorite. (In the inner zone hypersthene, in the exterior zone green hornblende and spinel.) Between the magnetite and the labradorite is a reaction rim consisting chiefly of light-brown hornblende. (28:1.)

ferromagnesian silicate. This boundary, 4 per cent Fe_3O_4, is, however, quite approximate, and it is to a great extent dependent upon the composition of the magma.

In gabbro rocks, rich in plagioclase, and in anorthosite gabbros, the boundary seems to lie somewhat higher, viz., according to

my earlier determination of the rock from Lofoten described above, at about 8–9 per cent magnetite.[1]

We here interject the remark that the simultaneous crystallization of "titanomagnetite" and hypersthene, diallage, or olivine, taking place at an early stage of crystallization in norites and gabbros (with olivine gabbros, etc.) which carry much ferromagnesian silicate and some few per cent of iron ore, explains how the well-known magmatic differentiation products of "titanomagnetite" in these rocks are characterized by an enrichment of "titanomagnetite" plus the ferromagnesian silicate in question, consequently hypersthene, diallage, or olivine respectively.

Magnetite in the granitic rocks.—Here the magnetite, when present in a quantity of at least 0.5 or 1 per cent, commences crystallizing at a very early stage, viz., a little later than the apatite, which, as is known, often appears as idiomorphic needles in the magnetite, but on the other hand earlier than the biotite or other ferromagnesian silicates which, in part, were deposited on the *Fixkörper* magnetite.

At this early stage, however, the whole quantity of magnetite was not crystallized. We are thus able to detect exceedingly small magnetite individuals in the groundmass, for example, in quartz porphyres; and the many analyses (Nos. 14*c*–29*c*) of glass, ground-mass, and intervening mass between the orbicules, etc., in granite without exception show a little Fe_2O_3 and FeO. In the granitic eutectic which at last results, there usually seems to appear about 1 per cent Fe_2O_3 and FeO, of which, however, a little is bound up with the small quantity of ferromagnesian silicate, and a little Fe_2O_3 probably also appears as a solid solution in the feldspar. The magnetite seems to appear in the final eutectic in a quantity of about 0.5 per cent.

For the magnetite as well as for the ferromagnesian silicates the solubility is much less in the granitic than in the gabbroidic magmas.

Spinel.—Magnetite and spinel (pleonast and hercynite) (Mg, Fe) (Al_2, Fe_2)O_4, with at most about 7–10 per cent Fe_2O_3, but sometimes with more FeO than MgO, form between them *a* series of

[1] Foslie (*loc. cit.*) sets the boundary still somewhat higher. This, however, is not in accordance with my investigations of the problem in hand.

discontinuous mix-crystals with a eutectic boundary consisting of about 2.5–3 per cent spinel:97.5–97 per cent magnetite (cf. my treatise, cited p. 2, "Über das Spinell:Magnetit-Eutektikum," 1910).

In the igneous rocks, containing *primary* spinel, this mineral, as is well known, belongs to the very first stage of crystallization when it is present in at least about 0.1 per cent by weight. But if the quantity of spinel sinks still lower the case is different. As an example we may mention that in the olivine-hyperite, illustrated in Figure 33, where first some olivine crystallized and later a little magnetite, some quite small individuals of spinel often appear in the magnetite. The spinel forms about 1 per cent of the magnetite, and since the latter forms 1–2 per cent of the entire rock, the quantity of spinel may consequently be set to 0.01–0.02 per cent. And this minimal quantity of spinel did not crystallize until mineral No. 1, in this case the olivine, had commenced forming.

Further we shall mention an ilmenite-norite from Storgangen near Soggendal (Ekersund), consisting of about 40 per cent ilmenite, 40 per cent hypersthene, 20 per cent labradorite, and in addition 1–2 per cent biotite, about 0.2 per cent pyrite, and about 0.1 per cent spinel. The latter is intensely green and must be termed hercynite.

The pyrite appears in small cubes (with edges 0.1–0.3 mm.) and on these quite small spinels (0.04–0.1 mm. large) have grown. These were deposited on the crystals of pyrite,[1] which in this manner served as *Fixkörper* in the still molten magma. The small spinels show an octahedron boundary against the surrounding, later mineral—ilmenite, labradorite, or hypersthene. FeS_2 as well as (Mg, Fe) $(Al_2, Fe_2)O_4$ were so little soluble in this magma that, although present in so small a quantity, they commenced crystallizing at an earlier stage than the silicates and the iron ore, but in such a manner that at the very first stage some pyrite solidified, and then, at somewhat lower temperature, the spinel.

The essential part of the *apatite* crystallizes, as is well known, at a very early stage. Sometimes, in the same thin section (Fig. 35), we may find a crystal of pyrite serving as *Fixkörper* for the deposit

[1] The same sequence of crystallization, spinel later than pyrite, I have mentioned also in an earlier treatise, see *Zeitschr. f. prakt. Geol.* (1900), p. 238, Fig. 39.

of apatite, and in other places a crystal of apatite serving as *Fixkörper* for a deposit of pyrite, showing that the two minerals were formed practically simultaneously. At the same weight (as 0.1–0.25 per cent) of apatite, pyrite, and spinel, the crystallization of apatite and pyrite seems to commence at a somewhat earlier stage than the crystallization of spinel.

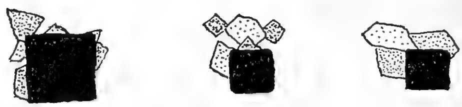

FIG. 34.—From an ilmenite-norite from Storgangen, near Soggendal, Norway. Three crystals of pyrite on which have been deposited small crystals of spinel, lying in ilmenite. (40:1.)

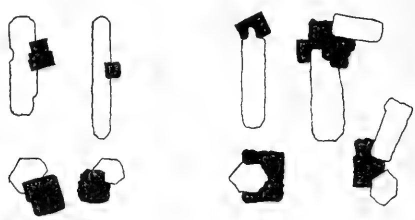

FIG. 35.—From a uralitic quartz-norite from Flaad, Evje, Norway. Black = pyrite, white = apatite. (40:1.)

PYRITE AND PYRRHOTITE

According to the precision-investigations by E. T. Allen, I. L. Crenshaw, and I. Johnson,[1] at the Geophysical Laboratory in Washington, the melting-points are for FeS, 1170±5°; for pyrrhotite, 1183° (to 1187°).

The analyses of phyrrhotite show, as is well known, a surplus of S above the proportion Fe:S, which the investigators mentioned (1912) interpreted to indicate that some S entered as a solid solution in FeS. Soon after, Docent C. W. Carstens,[2] of Trondhjem, pointed out that what enters in the solid solution must be FeS_2 and not S, an interpretation which, independently of Carstens' statement,

[1] *Amer. Jour. of Sci.*, Vol. XXXIII (1912).

[2] *Norsk geologisk tidsskrift*, Vol. III (1914).

was also later indicated by Posnjak, Allen, and Mervin[1] at the Laboratory in Washington.

That this is really so, appears theoretically through the fact that the process:

$$FeS+S \rightleftarrows FeS_2$$

is reversible. FeS in solid solution is able to absorb as much as 23 per cent FeS_2. The solubility of FeS in FeS_2 is, however, nil or minimal.

For the discontinuous binary mix-crystal system, only two cases exist, viz., either type V (with a eutectic) or type IV (with a bending-point). By the following observations we may determine to which of these two types the system $FeS:FeS_2$ belongs.

1. The pyrrhotite has a relatively low melting-point not only at the pressure of one atmosphere but, as appears from the statement given below, also at the high pressures prevailing in deep-seated magmas. The pyrite, on the other hand, crystallized in the deep-seated magmas at a very early stage and, consequently, at a relatively high temperature. The pyrite, accordingly, at high pressure may exist in the solid phase at a much higher temperature than the melting-point of the pyrrhotite. The genesis of the intrusive pyrite deposits proves that FeS_2 at very high pressure may also occur in the liquid phase. Pyrite consequently (at high pressure) has an even much higher melting-point than pyrrhotite.

2. Pyrrhotite (FeS with FeS_2 in solid solution, with as much as 23 per cent FeS_2:77 per cent FeS) has a higher melting-point (*G* on Fig. 36) than FeS. This already proves that the system belongs to type IV.[2]

3. We have a confirmation of this in the fact that in the ore deposits of pyrite and pyrrhotite, formed by magmatic differentiation, we always find, irrespective of the proportion by weight between the two sulphides, the sequence of crystallization, (1) pyrite (2) pyrrhotite, but, on the other hand, never the inverse sequence of crystallization, (1) pyrrhotite, and (2) pyrites. Regarding the

[1] *Amer. Jour. of Sci.*, Vol. XXXVI (1915).

[2] See my résumé-treatise, "Die Sulfid-Silikatschmelzlösungen" (1917).

resorption phenomena at the stage I to K we refer to the explanation in a later paragraph.

With regard to the structure of the deposits of *pyrrhotite-hypersthenites*, *-olivinites*, *-norites*, etc., formed by magmatic differentiation, I refer to my earlier publication "Die Sulfid-Silikat-schmelzlösungen" (in *Norsk geologisk tidsskrift*, 1917). I shall here only give a short summary.

We shall commence with the per se very rare *pyrrhotite-olivinites*, and as an example choose the deposit at Bruvand (Ballangen,

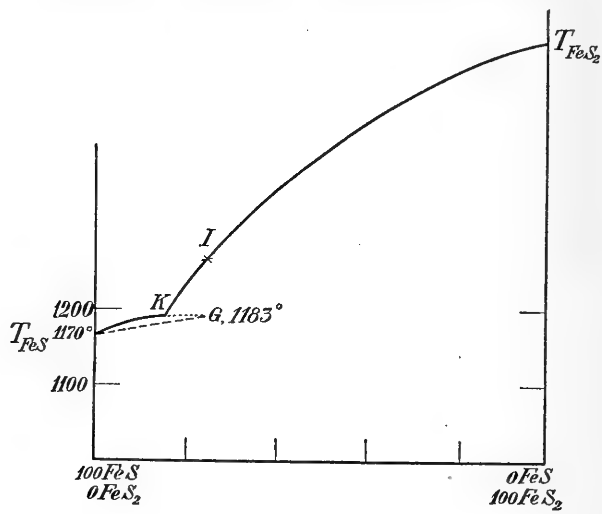

FIG. 36.—The binary system FeS: FeS, at high pressure. (The melting-points for FeS and pyrrhotite refer to low pressure.)

Ofoten in Norway) where the olivinite carries about 8 per cent sulphides, which we for brevity's sake shall term "nickel-pyrrhotite" (pyrrhotite with some pentlandite, chalcopyrite, and pyrite, in the present case with about 6–7 per cent nickel).

The rock consists, with deduction of the sulphides, of about 85–90 per cent olivine (according to chemical analysis $0.12\ Fe_2SiO_4 \cdot 0.88\ Mg_2SiO_4$, and accordingly quite poor in iron and not the least serpentinized), and in addition a little bronzite, primary hornblende, etc. The sulphides are not quite evenly distributed over the whole rock, but accumulated in small patches, especially on the boundary between the olivine grains (see Fig. 37).

This is most clearly shown in the thin sections (see Figs. 38 and 39), where the olivine individuals against the pyrrhotite show more or less well-developed idiomorphism; however, such that the edges are somewhat rounded.

We especially remark the quite good idiomorphism of the *small* olivine individuals against the pyrrhotite.

The rather common segregations of *pyrrhotite-hypersthenite* in hypersthene-rich norites carry from 10–20 to 50–60 per cent of nickel-pyrrhotite. Besides the sulphides, they may carry hypersthene alone or hypersthene accompanied by more or less plagioclase, etc. The hypersthene here appears as porphyritic crystals in the pyrrhotite; the boundary planes of the crystals, however, usually are somewhat rounded, with small bends, etc. (see Figs. 40–43).

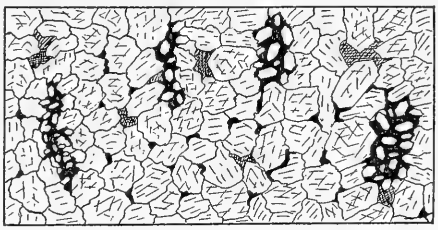

FIG. 37.—Pyrrhotite-olivinite from Bruvand, Ofoten, Norway. Light-structured mineral = olivine, dark-structured = bronzite and hornblende, black = pyrrhotite. (Natural size.)

A corresponding structure may also be observed in a *pyrrhotite-norite* from Dyrhaug in Værdalen, chiefly consisting of labradorite (Ab_1An_2, occasionally with zonal structure from about $Ab_{28}An_{72}$ in the kernel to about $Ab_{40}An_{60}$ in the exterior zone) and nickel-pyrrhotite (see Figs. 44, 45).

The structure here described can only be explained to show that the *olivine* (Figs. 37–39), the *hypersthene* (Figs. 40–43), and the *labradorite* (Figs. 44–45) *have crystallized at an earlier stage than the nickel-pyrrhotite.*

And this is in best conformity with the intervals of the melting-points, calculated for the different minerals at atmospheric pressure, and only quite unessentially changed at higher pressure (see a following paragraph).

The melting-point of pyrrhotite is 1183° (or 1187°), and for a sulphide mixture, consisting of predominant pyrrhotite, 3–6 per cent pyrite, some chalcopyrites, and some pentlandite, we may fix the crystallization interval from about 1190° (or about 1200°), for the commencement of crystallization of the pyrite, down to,

or perhaps a little below 1150° for the last remnant of sulphides, consisting of chalcopyrite with some pyrrhotite and some pentlandite.

FIG. 38.—Photomicrograph between crossed nicols (15:1).

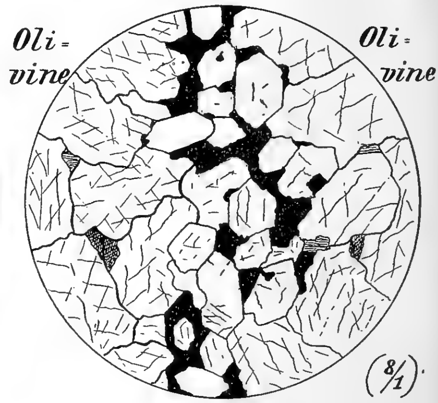

FIG. 39.—Drawing (8:1)

Pyrrhotite-olivinite from Bruvand (cf. Fig. 37). Two different portions.

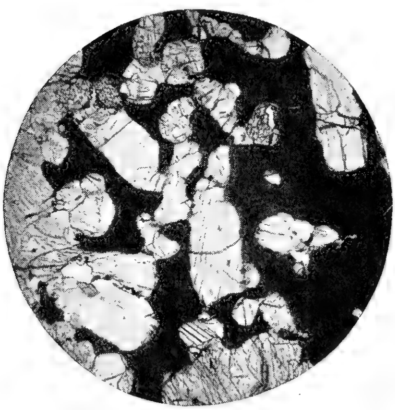

FIG. 40.—Photomicrograph (17:1)

FIG. 41.—Photomicrograph (17:1)

Pyrrhotite-hypersthenite from Romsaas (Fig. 40) and Messel, near Arendal (Fig. 41). In Figure 40 besides pyrrhotite there is only hypersthene. In Figure 41 besides hypersthene there is also a little labradorite (upper left). In Figure 41 two small holes in the thin section appear white.

For an olivinite, consisting of predominant olivine ($0.12\ Fe_2SiO_4 \cdot 0.88\ Mg_2SiO_4$) and a little orthorhombic pyroxene

(bronzite), likewise quite iron-poor, and iron-poor hornblende, we may reckon a crystallization interval from about 1500° down to 1400° or somewhat lower.

For hypersthene (with about $0.3FeSiO_3 \cdot 0.7\ MgSiO_3$) the crystallization interval, according to the earlier approximate determinations, lies at about 1300–1200°.

In a silicate mixture, consisting of predominant labradorite (Ab_1An_2) and a little diallage and hypersthene, the labradorite

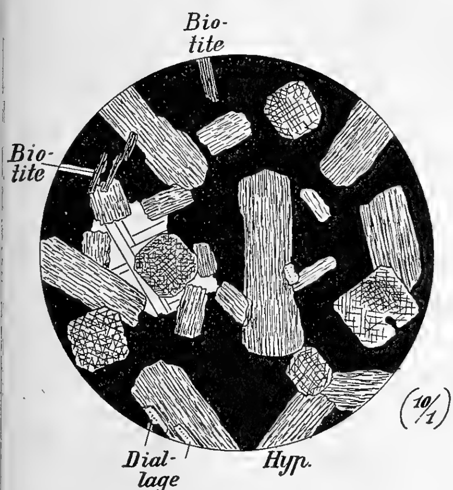

FIG. 42.—From Romsaas (10:1)

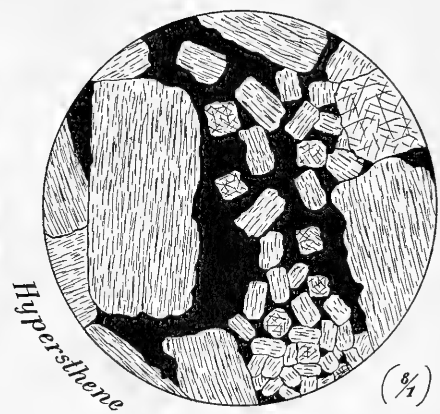

FIG. 43.—From Messel (8:1)

Drawings of pyrrhotite-hypersthenite. In Figure 42 besides hypersthene there is a crystal of labradorite (with twinning lamellae), a little biotite, and traces of diallage.

will commence crystallizing at about 1450°, and the essential part of the rock will be solidified at about 1300–1250°. The residual magma, present in quite small quantity, will crystallize along a eutectic boundary-line, down to about 1200° or probably a trifle lower. For the three pyrrhotite rocks just mentioned, the silicates must consequently in two cases have crystallized entirely and in the third case either entirely, or nearly entirely, while the sulphides were still in the liquid phase.

At temperatures of 1500°, 1400°, and 1300° melted silicate of the composition of olivinite, hypersthenite, or labradorite-rich

norite, may contain in solution only quite a small quantity of FeS (and still less of Cu_2S and NiS). Melted sulphide, consisting of FeS, etc., may, at the temperatures mentioned, dissolve only a trifle silicate.[1] The pyrrhotite-rich rocks mentioned must therefore during the interval of crystallization have consisted of *two liquids* (*two liquid phases*), *one silicate phase* with only a little dissolved sulphide and *one sulphide phase* without or with only a very inconsiderable amount of dissolved silicate.

FIG. 44.—Photomicrograph between crossed nicols (15:1).

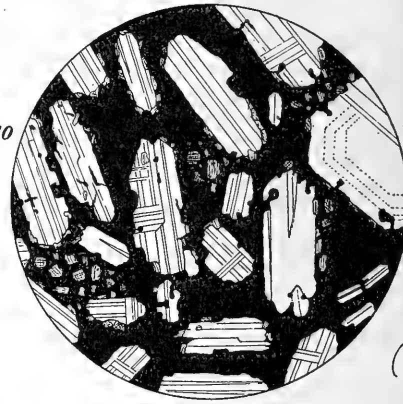

FIG. 45.—Drawing (8:1)

Pyrrhotite-norite from Dyrhaug, Skjækerdalen, Værdalen, Norway (cf. Figs. 20 and 21). Porphyritic labradorite in nickel-pyrrhotite, besides a little hypersthene, diallage, and biotite.

The common norites and gabbros usually contain 0.1–0.4 per cent S, which corresponds to about 0.25–1 per cent pyrrhotite, which, however, occasionally is accompanied by some pyrites. The pyrrhotite is here usually not evenly distributed in quite small individuals over the whole rock, but most often accumulated in somewhat larger lumps with a diameter of 0.5, 1, or 2 mm., occasionally more (see, for instance, Fig. 10).

Against the pyrrhotite the hypersthene as well as the diallage and labradorite here also appear with idiomorphic outlines. As an example we refer to Figures 12 and 21 and to Figure 46 from a

[1] See "Sulfid: Silikatschmelzlösungen," I (1919).

norite from Erteli, Norway, which, according to the analysis of the entire specimen, only contains 0.07 per cent S=0.18 per cent pyrrhotite, here and there accumulated in small lumps.

For the photomicrograph was chosen a locally pyrrhotite-rich part of the thin section, otherwise free from pyrrhotite. That pyrrhotite also crystallized last in common norites and gabbros is due to the circumstance that the intervals of the crystallization

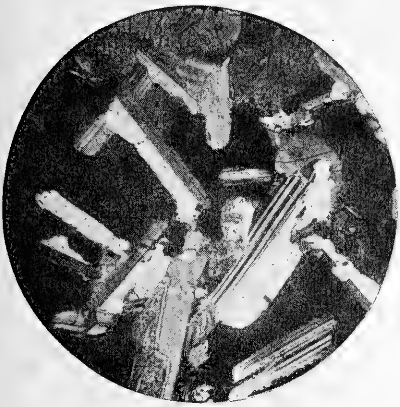

FIG. 46

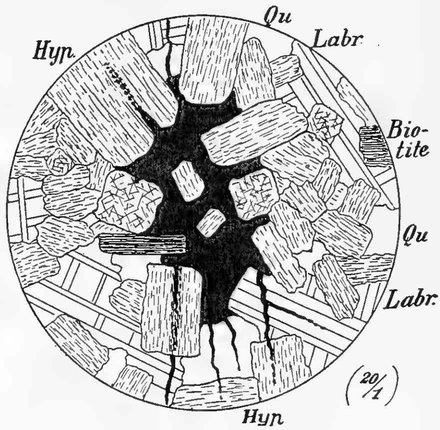

FIG. 47

FIG. 46.—Photomicrograph between crossed nicols (15:1). Pyrrhotite in norite from Erteli, from the same thin section as Figures 15 and 16. The rock contains 0.07 per cent S. Figure 46 is from a portion rich in pyrrhotite.

FIG. 47.—Quartz-norite from Romsaas, Norway (20:1). Illustrating a portion rich in pyrrhotite, from the same thin section as Figure 10. The hypersthene, biotite, and labradorite show idiomorphic contours against the pyrrhotite, and small veins of pyrrhotite cross the silicates.

for the silicates lay somewhat above, but for the pyrrhotite a little below, 1200°.

As well in pyrrhotite-hypersthenites and -norites, etc. with quite much pyrrhotite, as in the common norites, etc. with only about 0.2–0.5 or 1 per cent pyrrhotite, we frequently find *veins of pyrrhotite* branching off from relatively larger accumulations of this mineral, and intersecting the neighboring minerals.

We refer to Figure 47, where we see quite thin veins of pyrrhotite intersecting the hypersthene and the labradorite, which otherwise

show idiomorphic contours against the pyrrhotite. This phenomenon is due to the fact that the melted FeS even at *so low a temperature* as about 1200° is very thin fluid.[1]

FeS_2 is so little soluble in the silicate magma that the limit of the solubility, when only about 0.1–0.2 per cent FeS_2 is present, is usually reached by the cooling of the magma to somewhat above the upper boundary of the crystallization interval of the silicates. The melting-point of FeS_2 lies (at high pressure) at a still higher temperature. FeS_2 consequently, in the common igneous rocks, separates in the *solid* phase.

The solubility of *FeS* in silicate magmas at any temperature depends, as explained in my treatise "Die Sulfid-Silikatschmelzlösungen," on the chemical composition of the magma; the solubility under otherwise similar conditions increasing with the basicity. And as usual with dissolved substances, the solubility of FeS also increases with the temperature. In melts of gabbroidic composition, the solubility at 1350–1400° and at the pressure of one atmosphere is about 0.25–0.4 per cent FeS but decreases rapidly at a little lower temperature, as about 1300°. When the limit of solubility is reached by the cooling of the silicate magma, FeS accordingly separates in the *liquid* phase.

On account of the great difference existing at high pressure between the melting-points of FeS_2 and FeS, an essential difference arises with regard to the physics of the segregation: *FeS_2 crystallizes while FeS, on the other hand, separates in liquid phase* and remains in this condition till near the termination of the solidification of the rock.

The *intrusive deposits of pyrite*, which often are characterized by nearly exclusively pyrite, and which usually do not contain any pyrrhotite, prove also that FeS_2 at especially high pressure may exist in the liquid phase.

The formation of these deposits is probably due to the fact that locally in the igneous magmas so much FeS_2 has been present, that the latter at especially *high* temperature has been secreted at

[1] See "Die Sulfid-Silikatschmelzlösungen" (1917), where I have discussed the importance of the thinness of the pyrrhotite magma for the interpretation of the morphology of these deposits.

a stage above its melting-point, accordingly in the liquid phase. But the rocks normally contain so little FeS_2 that in the granites, syenites, etc., it does not secrete until below the melting-point of the mineral, consequently directly in the solid phase.

ON THE EARLY CRYSTALLIZATION OF "APATITE AND ORES"—OR OF THE "TELECHEMIC" MINERALS

All *phosphates* (apatite and monazite with zenotime in granite-pegmatite dikes), *sulphides* (pyrite, pyrrhotite, chalcopyrite, pentlandite, etc.), *zircon*, *corundum*, *spinel*, *chromite* (in peridotites), *hematite* (and ilmenite and magnetite in the acid igneous rocks), *titanates*, *tantalates*, and *niobates* (the last ones in granite-pegmatite dikes), further *carbon* (graphite, diamond) and native *metals* (nickel-iron, platinum), as is well known, belong to the very first products of crystallization[1] in the igneous rocks. All these substances occupy an exceptional position, being extremely little soluble in silicate magmas at a temperature somewhat above the beginning of crystallization of the silicates, and this independent of their melting-points (very high in corundum and spinel; medium-high for instance in fluorine-apatite, about 1650°; chlorine-apatite, about 1530°; hematite, about 1560°; and probably somewhat lower for instance in titanite, and for pyrrhotite only 1183°).

For all these *early crystallizing* minerals, which in *chemical respect diverge very considerably from the composition of the silicates*, I have in "Die Sulfid-Silikatschmelzlösungen" proposed the term "*telechemic*" minerals (τῆλε, *tele* = distant, the same root as in telegram, telephone, telepathy, etc.).

ON "REACTION RIMS"

We choose for an example the *coronation of olivine* bordering on plagioclase, described from oliviniferous gabbros, etc., by several earlier investigators, with an *inner* zone (adjoining the olivine) consisting of hypersthene, and an *outer* zone (adjoining the plagioclase) consisting of hornblende almost always associated with some amount of spinel. In addition there occurs exceptionally a third

[1] For the pyrrhotite we must change the term crystallization to segregation (to a special fluid phase).

zone, mainly consisting of garnet, immediately adjoining the plagioclase.

1. As is well known, these zones only occur on the boundary between the olivine and the plagioclase, but never between the olivine and the pyroxenes or the magnetite.

2. The zones occur not only in rocks rich in olivine in which the olivine has in a great measure crystallized earlier than the plagioclase (see, for instance, Fig. 27), but also in rocks deficient in olivine where the plagioclase, showing idiomorphic outlines against the olivine, must have crystallized earlier than the latter. I beg to refer to Figures 48–49, representing an olivine-hyperite from Elverum, Norway, consisting of *ca.* 65 per cent labradorite, *ca.* 10 per cent olivine, and *ca.* 25 per cent diallage, the labradorite having early crystallized in lath-shaped crystals, often projecting into the olivine which did not begin to form till a somewhat later stage. We fix on the fact that the coronation zones sometimes branch a little into the labradorite, in part following local transverse fissures, in part twin lamellae, and in part the boundary between various individuals of plagioclase. I further refer to the oliviniferous labradorite rock (with Ab_1An_1) illustrated by Figures 23–24, in which however, the coronation zones are so thin that they are not recognizable in the photomicrograph. High magnifying powers are here required to enable a close examination.

From the very fact that the coronation zones only occur between the olivine and the plagioclase, and are never found between the olivine and other minerals (diallage, hypersthene, etc.), it may be inferred that in rocks rich in olivine the metamorphosis (*Umbildung*) of the olivine did not take place at a conjuncture when only the olivine had crystallized while the rest were still in a molten state. The metamorphosis accordingly, as has been earlier pointed out, particularly by Frank D. Adams, must be a phenomenon belonging to the solid phase. This theory, moreover, is verified by the occurrence of the metamorphism of the olivine as well as of the plagioclase, attached to the *boundary plan* between the two minerals, not only where the olivine is to a great extent crystallized earlier than the plagioclase, but also where the plagioclase to a great extent crystallized earlier than the olivine.

3. The coronation occurs in rocks not dynamometamorphosed, and accordingly, as has been pointed out by earlier investigators, it is no function of any orogenic pressure.

4. The zones here treated I have observed in all of the very many microscopic thin sections that I have examined, of oliviniferous gabbros, norites, and anorthosites (with labradorite or still more basic plagioclase). In fact, the occurrence of the zones in these deep-seated rocks must be accepted as a general phenomenon.

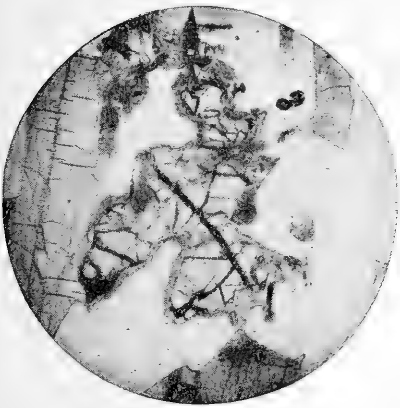

FIG. 48.—Photomicrograph (24:1)

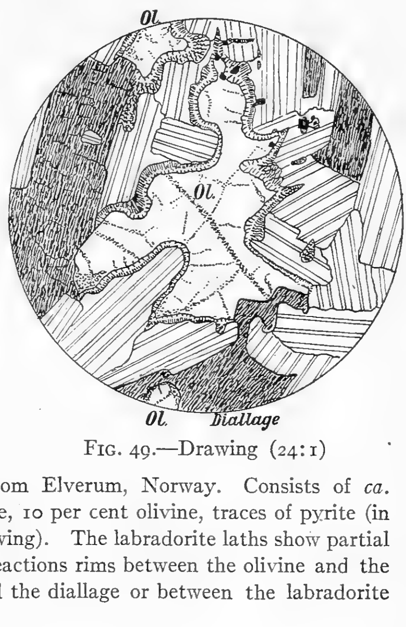

FIG. 49.—Drawing (24:1)

Hyperitic-structured olivine-gabbro from Elverum, Norway. Consists of *ca.* 65 per cent labradorite, 25 per cent diallage, 10 per cent olivine, traces of pyrite (in small cubes in the upper left portion of drawing). The labradorite laths show partial idiomorphism against the olivine (Ol.). Reactions rims between the olivine and the labradorite but not between the olivine and the diallage or between the labradorite and the diallage.

5. The thickness of the zones increases in the deep-seated rocks, as a general rule, proportionately to the amount of An contained in the plagioclase.

I cite: In the oliviniferous labradorite rock (Figs. 23–24) just mentioned (with Ab_1An_1) the total thickness of the zones regularly amounts to only 0.002 mm., sometimes to 0.005–0.010 mm., and exceptionally reaches 0.015 mm. In most olivine-hyperites (with plagioclase about Ab_1An_2), the thickness, as a general rule, is nowhere less than 0.06–0.08 mm.; most frequently it amounts to

0.10–0.12 mm., sometimes it reaches 0.15–0.20 mm., and quite exceptionally even rises to about 0.3–0.4 mm. In oliviniferous rocks with bytownite, the thickness increases to a still larger amount, and exceptionally, in deep-seated rocks containing much basic plagioclase, the whole quantity of olivine may even be spent in forming the zones.

6. The total chemical composition of the zones is equivalent to the chemical composition of olivine plus plagioclase.

It should be particularly emphasized that the zone next the olivine consists of hypersthene, which may be chemically considered as olivine less half the contents of $MgO+FeO$, whereas the zone next the plagioclase consists of hornblende to which in most cases is added some spinel, which may be accounted for by the composition of the plagioclase with some addition of $MgO+FeO$.

The coronation of the olivine against the plagioclase—or, as it may be quite properly expressed as well, the coronation of the plagioclase against the olivine—is a phenomenon belonging to *the solid phase of the minerals after finished crystallization.* My son, Th. Vogt, state geologist, has pointed out to me that to these contact-new-formations the common physicochemical laws concerning the *reaction between two solid phases* are fully applicable.[1]

This reaction is advanced by *high temperature*, and besides is a function of *time*. Therefore it must be assumed that the coronation began to take place immediately after the formation of the two minerals reacting on each other, and went on down to a certain limit of temperature (by way of example, 900°, 700°, 600°, or perhaps lower). The more slowly the cooling took place during the interval of reaction, the more the reaction was intensified. We accordingly always meet with reaction rims between olivine and basic or intermediate plagioclase in deep-seated rocks, but not—or to a much less degree—in the dike and effusive rocks which were more quickly cooled.

[1] See Beyschlag-Krusch-Vogt, *Erzlagerstättenkunde*, Vol. I (2d ed., 1914), p. 122. I beg to point out that the physicochemical laws for the reaction in the solid phase will no doubt throw quite a new light on numerous geologic processes, particularly on dynamometamorphism and contact-metamorphism. Upon these processes the influence of the varying relations of time, pressure, and temperature will particularly be felt.

The substances most easily soluble in acids generally have the greatest power of reaction. Therefore it is obvious that, as a general rule, the reaction rims should increase according to the amount of An contained in the plagioclase, as the solubility in acids of the plagioclase increases with its contents of An—and in plagioclase more basic than Ab_1An_1 the solubility increases obviously. Both olivine and basic plagioclase are much more easily soluble in acids than, for instance, the pyroxenes, the hornblendes, the micas, or the acid plagioclases. It is therefore easily accounted for that just between olivine and *basic* plagioclase the reaction rims are generally most strongly emphasized.

The general view here maintained of the reaction in the solid phase between olivine and plagioclase may in principle be transferred also on the corresponding reaction rims between diallage and basic plagioclase, between magnetite or titanomagnetite and plagioclase, etc.

[*To be continued*]

SUGGESTIONS AS TO THE DESCRIPTION AND NAMING OF SEDIMENTARY ROCKS

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It seems unnecessary to apologize for suggestions as to describing and naming sedimentary rocks.[1] Shaw has recently said:

> The need of carefully recorded descriptions of the physical characteristics of ancient sediments is especially worthy of emphasis. . . . . Notwithstanding the fact that advances have been made, there is as yet no adequate systematic classification that is generally acceptable. There is not even a satisfactory nomenclature.

To attempt a thoroughgoing logical classification of sedimentary rocks is beyond the scope of this article. Unless Grabau's compounding of prefixes into one dinosaurian term is deemed objectionable, his classification can scarcely be bettered. Here only two lines of suggestion are offered. One aims at systematizing field descriptions of rocks. The other deals with problems of nomenclature which chiefly arose during the writer's examination of three hundred specimens and eighty thin sections of the so-called "Deadwood" formation in the Bighorn Mountains, Wyoming.

FIELD DESCRIPTION OF SEDIMENTARY ROCKS

Lahee, in a warning against the padding of rock descriptions, remarks that every detail not germane to the immediate purpose of a given report should be omitted.[2] Were reports used only by persons interested in such immediate purposes, the comment would hold good. But if one tries to discuss the conditions of sedimentation in any given region, he is either forced to personal reconnoissance of adjacent regions or to dependence upon published reports of those regions. For such purposes the average descriptions of sediments are lamentably inadequate. Accordingly, there is here

[1] This article was written in April, 1920.

[2] F. H. Lahee, *Field Geology*, p. 450.

advocated a list of points in the description of sedimentary rocks which, it is believed, should never be omitted.

The list is as follows:

1. *Location.*—Nothing is more provocative of time-wasting than for a second observer to search for a vertical section made "on Wolf or Cow Creek," said creek being in a narrow canyon, 15 miles long. Various more accurate modes of statement readily suggest themselves.

2. *Name of rock.*—This, a matter determinable partly in the field, partly in the laboratory, is later discussed.

3. *Massiveness or stratification.*—This should include mention of the average thickness of beds; or, if there are alternations, reference to average thickness for each type.

4. *Color, fresh and weathered.*—Names for color naturally involve a personal equation. Simplicity is often needed. Admirable as are the descriptions of sediments in such a professional paper as No. 78, the amount of space devoted to refinements of color might well have been allotted to other details. Too often the color of pebbles in a conglomerate is not considered apart from the color of the matrix; secondary color, whether of pebbles or cement, is not distinguished from primary, etc.

5. *Hardness.*—The employment of a mathematical scale, as with minerals, seems impossible. Hardness, however, is usually a result of cementation. The following terms are perhaps usable: (*a*) soft: rock easily broken between the fingers; (*b*) subhard: rock breakable by a light hammer tap; (*c*) hard: rock broken only by a sharp hammer blow; (*d*) superhard: rock dense and resistant to the hammer. Doubtless, strength of arm is here a factor; but the personal equation is less than with color.

6. *Size of grain.*—In the field, measurements can rarely be made. Accordingly, exact limits for such terms as fine-grained, medium-grained, coarse-grained are not advocated. However, the term "pebble" should denote only material over a given size, say $\frac{1}{4}$ in. in diameter. The most inexpert assistant knows such terms as oölitic, pisolitic, etc.

7. *Mineral composition.*—In the field, statement of mineral composition is seldom more than a guess. The use of proper

qualifying adjectives, however, and attention to the order of their arrangement, would be valuable. Arenaceous glauconitic limestone should not be used when glauconitic arenaceous limestone is meant.

8. *Luster.*—Only occasionally does one find such remarks as "vitreous sediments," "clayey surface," etc. Perhaps luster has no great significance. Yet might not freshly exposed rocks be said to have vitreous, subvitreous, dull, earthy, etc., luster?

9. *Lateral variation.*—The importance of this factor can hardly be overemphasized. Any complete description of the "Fountain conglomerate" of the Front Range, e.g., would at once raise doubt as to the ordinary explanation of its origin.

It will be at once remarked that none of these characters throws so much light on the history of a formation as does, say, the mere mention of drying-cracks.[1] May that not be because drying-cracks have been studied by a geologic genius? The trouble with our field descriptions is that we incline to note only the unusual. Moreover, the extension of the writer's list to include drying-cracks, ripple marks, jointing, topographic result of erosion, and other matters, would defeat his very practical aim. An elaborate schedule cannot be recalled in all its ramifications. The simple list here advocated can be easily mastered.

THE NAMING OF SEDIMENTARY ROCKS IN GENERAL

As stated, no attempt is made to present a complete classification of sedimentary rocks. It seems possible, however, to group all sedimentary rocks as follows, partly by the aid of field observations, partly by microscopic examination:

A. ORGANIC—

1. Calcareous rocks
2. Ferruginous rocks
3. Siliceous rocks
4. Carbonaceous rocks
5. Rarer types

[1] There is a further educational value in detailed, systematic description. In courses in advanced general geology, students are often required to make vertical sections based on folios, etc., and even to describe formations in class. If descriptions are so meager that students cannot visualize the sections they describe or construct, is not the value of such work greatly vitiated?

B. Inorganic—
- 1. Of chemical origin
  - *a*) Calcareous rocks
  - *b*) Ferruginous rocks
  - *c*) Siliceous rocks
  - *d*) Halides
  - *e*) Sulphates
  - *f*) Rarer types
- 2. Of mechanical origin
  - *a*) Uncemented
  - *b*) Cemented

It is highly doubtful whether new names under all these headings are needed. Whoever is not enamored of the emptiness of Hegelian dichotomy and its dread of intercrossing rubrics will rest satisfied with the use under A of such stand-bys as bacterial limonite (A 2), diatomaceous earth (A 3), or coal (A 4). Like granite, diorite, gabbro, these names mean something to the geologist who cannot be forever searching his Greek or Latin lexicon for prefixes and suffixes. Nor can any inventor of systems hope to force aside such terms as travertine (B 1*a*), clay ironstone (B 1*b*), novaculite (B 1*c*), salt (B 1*d*), gypsum (B 1*e*).

THE NAMING OF ROCKS UNDER A 1 AND B 2

Obviously, if such familiar names as onyx, sinter, uintaite gypsum are not readily replaceable, there must be a deeper reason even than their establishment in usage. The analysis of this reason results, it is believed, in the discovery that, preponderantly, such names have three great values: (1) they are simple; (2) they aid clear visualization; (3) they suggest a dominant mode of origin. Dissatisfaction with such terms as limestone and sandstone probably arises, it may be unconsciously, from the failure of these names to meet the two latter criteria.

Does not, then, the renaming of organic calcareous rocks and of "clastics" involve, as its pivotal motive, the use of a terminology which shall at least approach the suggestiveness of "coal,"" tripoli," etc.? Furthermore, should not the new names be simple or at least readily comprehensible? Should they not aid visualization? And should they not somehow unveil the complex of conditions under which a given rock originated?

In pursuance of this belief, the renaming of rocks falling under A 1 and B 2 is considered from five points of view: (1) rock source of the sediment; (2) size and shape of grain; (3) degree of cementation; (4) mineral composition; (5) fossil content.

1. *Rock source.*—Seemingly, there is much satisfaction in speaking of aqueous and eolian sediments, perhaps even more in mentioning an anemopotamoclast. Nevertheless, such terms do not aid the sedimentary-rock student in the same way that peridotite or bostonite aids the igneous-rock student. True, one cannot assert that sands from gabbro rock sources will look different from sands from granite rock sources; of the requisite thin sections there are too few descriptions to tell. In the writer's own experience, nevertheless, Cambrian sands derived from the microcline granites of Wyoming do seem to have a characteristic appearance under the microscope. Would it not secure definiteness of description if sedimentary rocks, mainly, of course, sandstones, had prefixed to their colorless names such terms as granitogene, gabbrogene, quartzitogene?

2. *Size and shape of grain.*—Shape of grain should be an essential part of a microscopic description. The terms angular, near-angular, subrounded, rounded might constitute a useful series.

Size of grain is, under the microscope, capable of exact delimitation. The scale used in connection with igneous rocks is not, however, subdivided enough. Accordingly, the adoption of a modification of the New York City Aqueduct standard is advocated. The terms suggested chiefly, perhaps, apply to sandstones, somewhat to limestones, little to shales.

Sedimentary rock very coarse-grained	grains over 1 mm.
Sedimentary rock coarse-grained	grains between 0.5 and 1 mm.
Sedimentary rock medium-grained	grains between 0.25 and 0.5 mm.
Sedimentary rock fine-grained	grains between 0.1 and 0.25 mm.
Sedimentary rock very fine-grained	grains between 0.05 and 0.1 mm.
Sedimentary rock superfine-grained	grains below 0.05 mm.

By this scale, of course, arkoses, graywackes, conglomerates, breccias, even most grits would be very coarse-grained. It will be observed, however, that the terms in the table refer to measurements under the microscope and for sandstones, limestones, and

shales only. For field use in connection with arkose, graywacke, conglomerate, breccia, and grit, other sizings for fine-grained, medium-grained, coarse-grained are advocated.

Fine-grained arkose (graywacke, etc.)...........grains to $\frac{1}{8}$ in. in diameter
Medium-grained arkose (graywacke, etc.)....grains from $\frac{1}{8}$ to $\frac{1}{4}$ in. in diameter
Coarse-grained arkose (graywacke, etc.)........grains over $\frac{1}{4}$ in. in diameter

In most cases, particularly in conglomerates, maximum and minimum as well as average sizes should be noted.

Some term seems needed to denote a rock which is mainly an even-sized matrix, but contains a few pebbles over $\frac{1}{4}$ in. in diameter. Pebbled sandstone, pebbled limestone, pebbled shale are advanced.

3. *Degree of cementation.*—The microscope permits the abandonment of the field terms to denote hardness: soft, subsoft, hard, superhard. The following incomplete table is tentatively offered for criticism.

Rock Not Well Cemented (Primarily field terms)	Cemented; Grains Not Interlocked	Cemented; Grains Interlocked
Sandrock	Sandstone to quartzite sandstone	Orthoquartzite Paraquartzite
Limerock	Limestone	Orthomarble Paramarble
Magnesian limerock	Dolomite	Orthodolomite Paradolomite
Clay	Shale	Slate
Arkose	Arkosite to quartzite arkose	Arkositite
Glauconite sandrock	Glauconitite (existent ?)	Existent ?
Ferrite sandrock	Ferrite	Existent ?
Gravel	Conglomerate	Quartzite-conglomerate

Most of these terms are comprehensible at a glance. By orthoquartzite is meant rock cemented only through infiltration and pressure. By paraquartzite is meant quartzite mainly originating through contact metamorphism.

Doubtless it is illogical to remove quartzite, marble, and slate from the category of metamorphic rocks. However, quartzites, slates, and marbles are universally given a place in vertical sections and in geologic folios are described under sediments. Schists and gneisses are not so treated. Practical exigencies would seem to override Aristotelian "laws."

4. *Mineral composition.*—In the field, proportions between minerals cannot be determined. Under the microscope they can. The only true problems which arise concern the existence in given rocks of varying proportions of quartz and feldspar, of quartz and calcite, of quartz and glauconite, of calcite and siderite, of calcite and glauconite, of siderite and glauconite, or of various analogous but rarer combinations. At present, arenaceous limestone is used to denote a rock preponderantly calcareous and, may we say, one-fourth arenaceous. Calcareous sandstone is employed when such percentages are reversed. But in the writer's experience a number of rocks exhibit percentages of minerals close to 50:50. It is suggested that such terms as calarenite, sidarenite, glaucarenite would prove useful and not uneuphonious names for 50:50 combinations of calcite and quartz, siderite and quartz, glauconite and quartz. Limestone-glauconite, limestone-ferrite are examples of similarly compounded names for other mineral mixtures in rocks. Simple field names suggested on page 655 would thus be used only in emergencies.

The presence of glauconite involves a minor problem. Except when calcite is a cement, its presence in small percentage in a sandstone would hardly warrant the use of the name calcareous sandstone. Glauconite, however, throws some light upon the conditions of deposition. It is advocated that glauconitic as an adjective be employed even if the percentage in a sandstone or limestone be as low as 5 per cent.

5. *Fossil content.*—Fossiliferous sandstone, shale, and limestone are names already familiar. However, every geologist should recognize at sight the various invertebrate phyla and the main classes. And many geologists could thus characterize fossil-bearing rocks as predominantly graptolitic, coralline, vermicosic, pelmatozoic, bryozoan, brachiopodic, molluscan, trilobitic, etc. Coquina seems to be a term for pelecypodic limerock.

ILLUSTRATIVE DESCRIPTIONS OF SEDIMENTS

With some hesitation, the present discussion is closed with illustrative descriptions of hand specimens of Cambrian rocks and

of thin sections made therefrom. Each description, though brief, includes (1) locality, position in stratigraphic column, and description of field specimen; (2) texture, list of constituents, and relative proportion of more important minerals; (3) description of the chief mineral or minerals; (4) brief description of minor minerals; (5) name. Items (1), (2), (3), (5) are never omitted, and here each is given a separate paragraph; (4) is sometimes omitted. Terms logically connected are hyphenated, as fine-grained; glauconite-limestone; ferruginous-calcareous.

The descriptions are, as stated, purely illustrative. It may be added, however, that the writer has ventured partly to base upon these and other thin sections certain conclusions upon Middle Cambrian paleogeography rather at variance with the accepted account for Wyoming.

Wy 16: Taken in unnamed creek on south side of Duncom Mountain, $\frac{1}{8}$ mile east of Devil Canyon Road, 20 ft. above granite.

Description of hand specimen: Arkose, massive, gray-yellow to buff, weathering brownish-white to gray, with grayish-pink irregular lenses, cross-bedded; pebbles quartz and feldspar, usually about $\frac{1}{4}$ in., in finer but arkosic matrix; quartz angular to sub-rounded, iron-stained; feldspars pink angular cleavage fragments; traces of basic material?

Texture granular-fragmental. Constituents quartz, 80 per cent; microcline and a plagioclase near oligoclase, 15 per cent; small amounts of orthoclase, biotite, ilmenite, zircon, apatite, as accessories, and kaolin, sericite, leucoxene, muscovite, limonite as alteration products. Liquid and gas inclusions in quartz. Cement quartz and limonite.

Two marked groups of quartz grains. Larger average 0.4 to 0.5 mm. in diameter, smaller 0.06 to 0.08 mm. Larger grains sub-rounded to oval, smaller angular. Vein quartz suggested by wavy extinction.

Microcline fragments quadrangular, fresh; plagioclase same. Either rarely over 0.08 mm. Muscovite apparently not primary.

Classed: granitogene fine-grained arkose.

Wy 23: Taken at Middle Fork of Crazy Woman Creek,[1] 15 ft. above granite.

Description of hand specimen: Arkose, quartzitic, red-brown, massive, coarse, interbedded with finer-grained non-arkosic sandstones. Quartz grains rounded, uniform-sized, up to $\frac{1}{8}$ in. in diameter. Feldspars fresh cleavage fragments, about 33 per cent of rock, up to $\frac{1}{8}$ in. in diameter. General vitreous luster.

Texture granular-fragmental, secondary growth of quartz grains. Constituents quartz, 95 per cent; small amounts of biotite, limonite, apatite, zircon (?). Cement quartz and limonite.

One quartz fragment 3 mm., largest otherwise 0.15 mm.; 50 per cent of quartz 0.8 to 1 mm.; rounded and oval.

Classed: granitogène fine-grained arkosite.

Wy 27: Taken at Johnson Creek, 4 ft. above granite.

Description of hand specimen: Sandstone, massive, buff, with $\frac{1}{16}$ in. bands of chocolate-brown, doubtless iron stain along obscure bedding planes; contains rare pebbles of purplish shale and yellowish quartz, minimum $\frac{1}{8}$ in., maximum 1 in. In general medium-grained, friable, subvitreous luster.

Texture granular-fragmental. Constituents quartz, 60 to 80 per cent, dependent on amount of feldspar, now altered to sericite; small amounts of microcline, biotite, hornblende, plagioclase near oligoclase, zircon, apatite, magnetite, ilmenite (inclusion) as accessories, and kaolin and sericite as alteration products. Liquid and gas inclusions in quartz. Cement limonite, kaolin, sericite, possibly chalcedony.

Quartz vari-sized, largest grain 0.55 by 0.25 mm., no grains below 0.05 mm.; near-angular to subrounded, frequently oval. Feldspars suggest secondary growth, averaging 0.05 to 0.08 mm.

Classed: medium-grained pebbled sandstone.

Wy 50: Taken $\frac{1}{4}$ mi. from mouth of tributary, flowing westward from Hunt Mountain to South Beaver Creek, in float not far above granite; similar material in place, 30 ft. south and 10 ft. above granite.

[1] The belt of exposure is so narrow that this locality seems sufficiently identified.

Description of hand specimen: Arkose conglomerate, pebbles not ranging above $\frac{3}{8}$ in. Prevailing color where fresh dark-brown shot with gray, the pink of feldspar sharply contrasting; from dirty grays and whites of weathered surface pebbles stand out in relief. Fucoid markings on what may be bedding-planes 1 in. apart. Pebbles 90 per cent quartz, faintly brownish-green, rounded to near-angular, breaking with matrix; feldspars pink fresh cleavage fragments. Matrix 90 per cent of rock, sand, fine-grained, dull to earthy luster.

Texture granular-fragmental, large grains showing micrographic intergrowth. Constituents quartz 90 per cent; microcline, orthoclase, uncertain plagioclase 8 per cent; small amounts of biotite, apatite, zircon as accessories, and sericite, kaolin, limonite, and chlorite as alteration products. Liquid and gas inclusions in quartz. Cement a sericitic-kaolinic-limonitic "mess."

Quartz vari-sized, 1.5 by 0.8 mm. in larger grains, perhaps vein quartz, to judge by wavy extinction; average grains, 0.08 mm.; rounded to near-angular. Microcline fresh, 0.03 to 0.04 mm. Orthoclase same size, largely sericitized. Much organic material, seemingly chitinous.

Classed: ferruginous arkose-conglomerate.

Wy 93: Taken on Willow Creek at Burgess Ranger Station, 2 ft. above granite.

Description of hand specimen: Shale, thin-bedded, green when fresh, sparsely specked with glistening mica flakes and containing lenses of whitish-green sandstone, 1 in. long; on weathered surface bluish-black. Hard, arenaceous, fracturing irregularly, fresh surface of dull luster, bedding-planes subvitreous luster and slightly wavy.

Texture granular-fragmental, pilitic through alteration and with parallel arrangement of minerals, excluding quartz. Constituents chlorite, sericite, epidote, presumably alterations from biotite, muscovite, feldspar; small amounts of quartz, plagioclase, magnetite, zircon. Glauconite indeterminable. Cement a sericitic-chloritic felt.

Quartz rarely 0.06 mm., very angular. Plagioclase rare. Biotite bleached.

Classed: micaceous-arenaceous shale.

Wy 2: Same locality as *Wy 16*, 40 ft. above granite.

Description of hand specimen: Sandstone, thin-bedded, alternating and interleaved with fissile green shales; now banded green and white, subsoft, micaceous, now reddish-green, medium-grained, calcareous.

Texture granular-fragmental. Constituents quartz 90 per cent; glauconite 8 per cent; magnetite, ilmenite, calcite as accessories, limonite and leucoxene as alteration products. Cement calcite.

Quartz averages 0.05 mm., angular; rare vein quartz. Glauconite in aggregates (0.16 mm. diameter) oval to rounded, bordered and veined by limonite; no seeming relation to shell fragments.

Classed: glauconitic superfine-grained sandstone.

Wy 9: Same locality as *Wy 16*, 170 ft. above granite.

Descripton of hand specimen: Glauconite sand, massive, emerald green, crumbling, coarse-grained, subvitreous luster.

Texture granular-fragmental. Constituents glauconite 95 per cent; limonite as alteration product 5 per cent. Cement limonite. Various chitinous fragments and rods.

Glauconite aggregates as in *Wy 2*, but fresher.

Classed: glauconite sandrock.

Wy 25: Same locality as *Wy 27*, approximately 30 ft. above granite.

Description of hand specimen: Sandstone, massive, buff where fresh, weathering dark-red. Superhard, fine-grained, dull luster. Weathering exhibits buff nodules in relief.

Texture granular-fragmental. Constituents quartz 85 per cent; small amounts of biotite, muscovite, glauconite, magnetite, ilmenite, plagioclase, microcline, zircon, apatite as accessories, and leucoxene, kaolin, limonite, sericite, chlorite, epidote as alteration products. Liquid and gas inclusions in quartz. Cement quartz and limonite.

Quartz averages 0.2 mm., angular; fragments often elongate; one grain 0.6 by 0.35 mm; vein quartz rare. Plagioclase a mass

of sericite needles, microcline much fresher; average size for both, 0.05 mm. Biotite shredded. Muscovite largely secondary, but some long twisted primary fibers.

Classed: granitogene glauconitic-ferruginous fine-grained sandstone.

Wy 58: Taken at Turkey Creek, $\frac{1}{4}$ mi. south of Steamboat Point, 9 ft. above granite.

Description of hand specimen: Limestone, 1-in. lenses, reddish-brown, coarse-grained; occurring in shale, fissile, paper-thin, green with silky luster, trilobitic on bedding-planes (*Ptychoparia*), brachiopods rare.

Texture granular-fragmental. Constituents calcite 95 per cent; small amounts of quartz, magnetite, zircon, siderite as accessories. Cement calcite and a trifle limonite.

Calcite averages 0.75 mm.; often in rods and rectangular blocks, obviously fossil fragments, the remaining calcite due to solution and redeposition. Within shell fragments a finely comminuted mixture of quartz, calcite, siderite; limonite, probably from siderite, outlines fragment edges. Quartz 0.02 to 0.03 mm., angular.

Classed: coarse-grained brachiopodic limestone.

Wy 99: Taken at Deer Creek, 1 mi. northwest of Sheep Mountain, 100 ft. below the persistent sandstone described as *Wy 89*.

Description of hand specimen: Limestone, massive 1-in. bed between thick green shales; gray-blue, weathers dirty-brown, crossed by veins of calcite; hard, medium-grained, subvitreous luster, brachiopodic and trilobitic.

Texture granular-fragmental. Constituents calcite 90 per cent; small amounts of quartz, magnetite, glauconite, pyrite, zircon, ilmenite as accessories, and leucoxene as alteration product.

Calcite 0.35 mm., anhedral; at times within shell fragments and then comminuted; "rods" clearly from genal spines of trilobites, fragments 4 to 5 mm. long. Quartz mainly in shells, angular, seldom as much as 0.04 mm. Glauconite seemingly developed in shells. Much chitinous material.

Classed: medium-grained brachiopodic trilobitic glauconitic limestone.

Wy 105: Taken at Horse Creek No. 3, 1 mi. west of Sheep Mountain, 125 ft. above granite and at much the same horizon as *Wy 99*.

Description of hand specimen: Limestone, massive, light-green and specked with hard black grains, weathering dirty brown-red. Hard, fine-grained, subvitreous luster, fractures conchoidally.

Texture granular-fragmental. Constituents calcite 60 per cent; glauconite 30 per cent; quartz 10 per cent. Magnetite and zircon as accessories, and limonite as alteration product.

Calcite 0.15 mm., probably due to recrystallization.

Calcite in shell fragments as in *Wy 99*. Quartz grains surprisingly large, average 0.1 mm. and up to 0.5 mm.; very angular. Glauconite in large aggregates, even 2 by 8 mm.; some curved, as if by replacement of whole fossils; indifferently near to or remote from magnetite; fresh. Limonite mainly from magnetite.

Classed: fine-grained brachiopodic trilobitic glauconite-limestone.

Wy 89: Taken from massive bench on south side of Bald Mountain.

Description of hand specimen: Sandstone, massive, pinkish buff, slightly splotched with brown, weathering dull grayish-white; subsoft, fine- to medium-grained, traces of shale streaks, earthy to subvitreous luster, brown splotches interpreted as oxidation of trilobite shields. Cross-bedded? Most persistent bed in the Cambrian below the flat pebble.

Texture granular-fragmented. Constituents quartz 95 per cent; small amounts of biotite, magnetite, ilmenite, plagioclase, glauconite as accessories, and leucoxene and limonite as alteration products. Cement quartz and calcite.

Quartz averages 0.15 mm.; largest grain 0.3 by 0.15 mm.; angular, sufficiently cemented to suggest quartzite; few indications of secondary growth. Grains limonite-rimmed. Biotite much shredded. Chitinous rods.

Classed: fine-grained quartzite-sandstone.

Wy 63: Same locality as *Wy 58*, 150 ft. above granite.

Description of hand specimen: Limestone, massive, gray-green, weathering reddish-brown to gray. Subsoft, medium-grained, argillaceous, arenaceous, glauconitic, ferruginous.

Texture granular-fragmental. Constituents calcite 50 per cent; quartz $33\frac{1}{3}$ per cent; siderite 10 per cent; small amounts of magnetite, ilmenite, zircon, muscovite, glauconite as accessories, and limonite as alteration product.

Calcite 0.2 mm., giving evidence of recrystallization; also as rods. Glauconite aggregates about 0.5 mm., much replaced by limonite. Limonite also edges siderite rhombs. Quartz 0.04 mm., near-angular to subrounded.

Classed: arenaceous trilobitic medium-grained glauconite-limestone.

Wy 13: Same locality as *Wy 16*, 260 ft. above granite.

Description of hand specimen: Limestone, thin-bedded, blue-gray, subsoft but brittle, the 1-in. layers separated by micaceous shale; subvitreous luster, black-specked, seemingly unfossiliferous.

Texture granular-fragmental. Constituents calcite 50 per cent; quartz 25 per cent; glauconite 20 per cent; small amounts of ilmenite, zircon, apatite as accessories. Liquid and gas inclusions in quartz. Cement calcite.

Quartz 0.1 mm., angular. Calcite grains 0.5 mm.; main occurrence as rods.

Classed: arenaceous trilobitic medium-grained glauconite-limestone.

Wy 87: Same locality as *Wy 89*, 4 ft. above that horizon.

Description of hand specimen: Sandstone, massive, gray-green with irregular-bedded effect due to light-colored stretches between dark-brown bands, with parallel orientation of *Dicellomus* shells; soft, coarse-grained, calcareous, particularly toward top, glauconitic, dull to subvitreous luster. Cross-bedded.

Texture granular-fragmental. Constituents quartz 80 per cent; glauconite 10 per cent; calcite 5 per cent; small amounts of biotite, muscovite (?), apatite as accessories. Cement calcite. Liquid and gas inclusions in quartz.

Quartz 0.2 to 0.5 mm., subrounded to oval, at times subhexagonal. Glauconite seldom related to shell interiors, not even in an admirable cross-section with quartz fragments in shell and cemented by calcite and limonite.

Classed: glauconitic medium-grained sandstone.

Wy 84: Taken at Cambrian Creek, tributary to East Fork of Little Bighorn River, long. 10° 45′ W., lat. 44° 50′ N., 34 ft. below the flat-pebble conglomerate.

Description of hand specimen: Limestone, massive, gray-green, weathering gray to reddish brown, subsoft, coarse-grained, arenaceous, glauconitic, subvitreous luster, with calcite veins and 1-in. crystals; breaks in smooth angular blocks; presents corrugated surface where calcite has dissolved on weathering.

Texture granular-fragmental. Constituents quartz 47 per cent; calcite 47 per cent; small amount of glauconite as accessory, and limonite as alteration product. Cement calcite.

Quartz averaging 0.3 mm., largest grain 0.5 by 1.8 mm.; subrounded; grains broken and healed by calcite; slight traces of secondary growth; small grains seemingly fragments of larger ones cemented. Calcite shows recrystallization. Glauconite aggregates about 0.1 mm., rounded. Bryozoan-like fragments.

Classed: glauconitic trilobitic medium-grained calarenite.

Wy 66: Taken on south side Tongue River, directly opposite mouth of Sheep Creek, 60 ft. below the Cambrian-Ordovician contact.

Description of hand specimen: Limestone, $\frac{1}{16}$- to 1-in. beds, greenish-white to buff, hard, fine-grained, argillaceous, arenaceous, subvitreous luster, slightly ripple-marked, raindrop-pitted (?).

Texture granular-fragmental. Constituents siderite 30 per cent; calcite 30 per cent; quartz 30 per cent; glauconite 5 per cent; a little apatite, ilmenite, muscovite, magnetite, and plagioclase; muscovite and plagioclase very rare; a little limonite as alteration product.

Calcite averages 0.1 mm., larger grains 0.2 mm. Quartz 0.05 mm., near-angular to subrounded; inclusions of hematite scales (?). No fossils.

Classed: glauconitic-arenaceous medium-grained siderocalcite.

Wy 72: Taken on East Fork of Little Bighorn River, 2 mi. northeast of Little Bald Mountain, at base of Ordovician "Bighorn dolomite."

Description of hand specimen: Limestone, 4- to 6-in. beds, gray-green, weathering light buff with yellow stains, hard, fine-grained, slightly dolomitic, argillaceous, dull luster, lower beds much jointed, causing weathering in subquadrate slabs; flat-spired gastropods abundant.

Texture granular-fragmental. Constituents calcite 99 per cent; a little accessory glauconite, limonite as alteration product. Cement calcite.

Calcite grains often recrystallized, largest 0.3 mm. and apparently cavity-filling.

Classed: gastropodic medium-grained limestone; traces of glauconite.

Wy 69: Same locality as *Wy 2*, 2 ft. below Cambrian-Ordovician-contact.

Description of hand specimen: Conglomerate, massive, greenish-gray; pebbles limestone, distinguishable with difficulty on weathered surface, greenish, subsoft, fine-grained, glauconitic, flattened, elongated, length $\frac{1}{4}$ in. to 2 in., often loose ochreous earth, lining cavities, when fresh breaking with matrix, and usually aligned in parallel planes, constituting 50 per cent of rock; matrix limestone, greenish, fine-grained, dull luster. (This is the famous flat-pebble conglomerate of Dakota and Wyoming.)

Texture conglomeratic. Constituents calcite 90 per cent; small amounts of glauconite, quartz, pyrite, magnetite as accessories, and hematite and limonite as alteration products. Cement calcite.

Calcite either as pebbles, merely fragmentary in slide, or as interlocked crystals in matrix. Pebbles characterized by calcite, criss-crossed and specked with glauconite (percentage from 25 to $33\frac{1}{3}$), and interlocked with quartz grains below 0.01 mm. diameter. Quartz very rare in matrix. Pyrite altering to hematite and limonite.

Classed: glauconitic flat-pebble limestone-conglomerate.

Wy 24: Same locality as *Wy 23*, 20 ft. below Cambrian-Ordovician contact.

Description of hand specimen: Seemingly *sandstone* (and so described by one observer), massive, pink, weathering fainter pink.

Superhard, fine-grained, argillaceous, and seemingly *slightly calcareous*, subvitreous luster.

Texture granular. Constituents dolomite 90 per cent; quartz 5 per cent; small amounts of magnetite, glauconite, ilmenite as accessories, and limonite as alteration product. Cement dolomite and limonite.

Dolomite in rhombs, inclusions of limonite; averages 0.07 mm. Quartz rarely over 0.02 mm., very angular. Glauconite altering to limonite, latter also between dolomite crystals.

Classed: glauconitic superfine-grained dolomite.

Reviews

The Cost of Mining. By James R. Finlay. Third edition (entirely revised, enlarged, and reset). McGraw-Hill Book Co., 1920.

The new edition of this standard work on mining costs in addition to amplifying and bringing up to date the data on mining costs found in the earlier editions, contains a considerable amount of material of broader economic interest relating to mineral resources. Chapter I, for example, discusses mineral wealth as a source of national power, chapter III treats of the nature and use of capital.

The cost of mining data is presented seriatim by mineral commodities and covers coal, iron, copper, lead, silver-lead, zinc, gold, and silver. The chapter dealing with each of these is commonly prefaced by some general discussion and by statistics of production. Cost data for iron-mining relate only to the Lake Superior region.

In the chapters devoted to copper occur such paragraph headings as "Geologic Unconformities at Jerome," "Characteristics of Belt Rocks," "Theories of Formation of Jerome Deposits," etc.; the book is therefore somewhat broader in scope than its title would suggest. The book commends itself not only to the engineer but to the economist, geologist, or geographer concerned in the rôle of mineral resources in the industrial life of the United States.

E. S. Bastin

Extracts from "The Mining Handbook," Geological Survey of Western Australia, Memoir No. 1, 1919. A series of advance separates of chapters from the foregoing *Handbook.*

This mining handbook is a worthy attempt to furnish to those interested in mining in Western Australia a large amount of varied information likely to prove of service to them in the exploitation of mineral deposits. The handbook includes chapters on the relations of physiography and of petrology to the exploitation of mineral deposits, chapters expounding the mining regulations and explaining various methods of governmental assistance to prospecting and mining. Then follow chapters dealing with the major base metals, with the various

steel-alloying metals, with the minor metals, and, finally, with coal and a few non-metallic mineral resources.

The chapter on physiography in its relation to prospecting and mining is mainly a discussion of the influence of topography on the discovery and development of ore bodies and, reciprocally, the influence of ore bodies on topography.

The chapter on minerals of economic value lists the composition and the principal physical properties of economic minerals and cites their main utilizations.

The chapter on petrology and its application in industry is an exposition of petrology in its simplest form, defining the principal rock-forming mineral and the principal groups of igneous, sedimentary, and metamorphic rocks and expounding the application of petrology in geologic surveying, in the study of ore deposits, in engineering, architecture, and agriculture.

The chapter on the relation of the law to prospecting and mining covers the legal restrictions governing the location and development of mineral deposits in Western Australia. Three points of contrast between the Western Australian mining laws and those of the United States are noteworthy. In the United States, discovery must precede the staking out of mining claims and ground cannot be validly held until there has been an actual discovery of mineral. In Western Australia ground can be marked out and held, even though no minerals have been discovered. In the United States title in fee simple to a mining claim is acquired by patent, subject to extra-lateral rights of adjoining claim-owners. In Western Australia the crown does not part with the title to the land. Leasehold is the rule, coupled with labor conditions.

In Western Australia the principle of extra-lateral rights, which has resulted in so much troublesome litigation in the United States, does not apply but the holder of a mining lease is only entitled to such portions of the lode or lodes as occur within the boundaries of his lease extended vertically downward from the surface.

An interesting feature of the Western Australian mining law is the provision for a reward of up to one thousand pounds, offered for the discovery of payable gold at a place distant more than two miles from any place where payable gold has up to then been discovered. Several other forms of governmental assistance to mining include advances for the purpose of pioneer mining and prospecting, the establishment and subsidizing of plants for ore treatment, assistance for drilling, including the purchase or hire of drilling plants, the advancement of money for

drainage, shaft sinking, and for development of transportation facilities to assist mining operations. A separate chapter, entitled "Assistance to Prospecting and Mining," explains these various forms of governmental assistance in detail and also lists geographically and by minerals all available government reports and maps covering mining districts. Another chapter is a glossary of common terms in mining and geology.

Among the major metals, iron ores though widely distributed in Western Australia have as yet been developed only on a small scale for the production of flux for copper and lead smelting and no detailed geologic surveys have been made of any of the iron deposits. Copper deposits, though widely distributed, have been developed only to a minor degree. The production of lead ores has been small.

Among the steel hardening metals, the production of manganese, tungsten, and molybdenum has been so small as to be essentially negligible. Among the rare metals, there has been little or no development.

The small tin production has come mainly from alluvial deposits, but in the Wodgina tin field, tin and tantalum occur in pegmatite dikes which, together with granite, intrude metamorphic sedimentary rocks. The chief constituents of these pegmatitic dikes are albite and quartz, with occasionally scaly lepidolite and tourmaline; in addition, orthoclase, mangano-tantalite, and tin occur in varying quantities, as well as some of the rare radioactive minerals. In the vicinity of and along the margin of many of the pegmatite dikes are bands and bunches of tourmaline, sometimes to such an extent as to make up fully one-third of the entire rock. One of the most conspicuous of the pegmatite veins, about half a mile in length and 30 feet in width, has proved to be sufficiently rich in tin and tantalum to be worked.

The tin ore, cassiterite, is concentrated along certain lines in the pegmatites and does not appear to be generally disseminated in minute quantities throughout its mass. The tin occurs in all shapes, from minute grains up to pieces weighing as much as 100 pounds.

The coal deposits of Western Australia range in age from Carboniferous, through Permo-Carboniferous to Mesozoic, Tertiary, and post-Tertiary. The only deposits which have been extensively mined are those of the Collie field of Permo-Carboniferous age. In this field the total thickness of the coal seams is about 137 feet. The coals are semi-bituminous, non-coking coals which are dirty to handle and deficient in volatile materials. It is interesting to note that the coals appear to be mainly of drift origin and to have been deposited by current action on an extensive basin or river valley. The banded appearance of most

of the coals and their relatively high percentage of ash is probably a result from this mode of origin as is also the absence of fire clays beneath them. The available reserves of the field are estimated at three and one-half billion tons.

E. S. BASTIN

Summary Report, Canadian Geological Survey. Ottawa, 1919. *Part C. Alberta-Saskatchewan Region.* Pp. 52.

1. "Cretaceous, Lower Smoky River, Alberta." By F. H. McLEARN.
2. "Geology of the Swan Hills in Lesser Slave Lake District, Alberta." By JOHN A. ALLEN.
3. "Northern Part of Crowsnest Coal Field, Alberta." By BRUCE ROSE.
4. "Gasoline in Natural Gas. Experiments on Alberta Gas." By D. B. DOWLING.
5. "Surface Deposits of Southeastern Saskatchewan." By J. STANSFIELD.

The annual *Summary Report of the Canadian Geological Survey* for 1917 and since is issued in parts and each part is designated by a letter of the alphabet. Before 1917 the whole annual *Summary* was bound in one large volume.

1. The Cretaceous begins with the Lower Cretaceous and extends into the Montana group of the Upper. Marine and non-marine formations alternate and the total thickness represented is about 4,470 ft. The Dakota sandstone cannot be recognized in its normal subaerial development. The beds dip to the south from 12 to 60 ft. per mile and represent the north limb of a very broad, shallow syncline. The structure is not favorable for oil.

2. The Swan Hills lie south of Lesser Slave Lake, have a maximum elevation of 4,320 ft. above sea-level, and represent remnants of a once more extensive, maturely dissected upland. The Cretaceous is represented by the Montana group. The basal member is marine, and the upper two members, the Sawbridge and Edmonton, are of fresh-water origin. The early Tertiary is represented by the Paskapoo formation but there is no marked unconformity between the Upper Cretaceous and the Tertiary.

3. Formations ranging from Devono-Carboniferous to Upper Cretaceous, probably Tertiary, in age, are described. Coal seams of

economic importance are found only in the Kootenay, and a large reserve of bituminous coal occurs within the Rocky Mountains.

4. This is a description of the apparatus and the results of a number of experiments carried on at various gas wells. At the pressures under which the tests were made the amount of gasoline in the Alberta gases per 1,000 cu. ft. varied from 0.1 pints to 3.7 pints.

5. This area is covered with glacial drift, averaging from 40 to 70 ft. in thickness. Two main terminal moraines cross the area but ground moraine covers most of the area. The residual alkali material formed in the dried-up sloughs contains only a very small per cent of potash, and is of no economic importance. The waters from the drift are hard and contain calcium and magnesium carbonates and sulphates while the waters from wells reaching the Tertiary strata are soft and contain considerable sodium chloride.

Part D. Manitoba Region. Pp. 19.

1. "Athapapuskow Lake District, Manitoba." By E. L. BRUCE.
2. "The District Lying between Reed Lake and Elbow Lake, Manitoba." By E. L. BRUCE.
3. "Reed-File Lakes Area, Manitoba." By F. J. Alcock.
4. "Wekusko Lake Area, Manitoba." By F. J. Alcock.
5. "Superficial Deposits and Soils of Winnepegosis Area, Manitoba." By W. A. JOHNSTON.
6. "Gold-Quartz Veins and Scheelite Deposits of Southeastern Manitoba." By E. L. BRUCE.

1. Chalcopyrite was discovered along joint or fracture zones in fine-grained, massive greenstone. Some distance from these occurrences the greenstone is intruded by granite and these deposits are directly related to these intrusions. With the present conditions of transportation mining conditions are not favorable for this area.

2. The geology of this area is much simpler than that of other nearby areas in northern Manitoba as the pre-Cambrian is represented by the Amisk series of greenstones and derived schists, and intrusive granites. The younger pre-Cambrian formations are absent, and since the crest of a large anticline crosses this area these younger formations have probably been removed by erosion from the crest of this anticline. Ordovician dolomites and Glacial and Recent deposits are noted. No

economic deposits have yet been discovered and on the whole conditions for the formation of ore deposits have not been as favorable as in nearby areas.

3. The pre-Cambrian rocks of this area are divided into an igneous complex consisting of altered volcanic and intrusive rocks, a sedimentary complex of granite-gneiss and staurolite-schist, and batholithic intrusives. Ordovician dolomites occur and Pleistocene and Recent deposits are abundant.

4. The geology of this area is very similiar to that of the Reed-File lakes area. A number of productive gold-bearing quartz veins occur near the borders of the granite masses.

5. Because of the practical exhaustion of homestead prairie land in easily accessible areas, a map of an area of about 1,500 sq. mi. around Lake Winnipegosis was prepared. This map will show the character of the soil and forests and will also indicate the land that can be readily cleared.

6. The gold-quartz veins in the pre-Cambrian rocks of southeastern Manitoba were sampled and assayed for both gold and platinum. Most of the assays showed a very small amount of gold present but in no case was platinum detected. In a fine-grained, massive, roughly sheeted, hornblendic rock scheelite occurs in small vuggy lenses not in all cases parallel to the sheeting. The returns from a shipment of the ore to the Ore Dressing Laboratory, Ottawa, were not encouraging.

Part F. Maritime Province Region. Pp. 36, figs. 3.

1. "Investigations in Western Nova Scotia." By E. R. Faribault.
2. "Investigations in Western Nova Scotia and New Brunswick." By Albert O. Hayes.
3. "Peat Investigations." By A. Anrep.

1. A description of a number of small manganese deposits in Nova Scotia and notes on the occurrence of platinum in the scheelite and gold veins of the gold-bearing series.

2. The drift over the Carboniferous rocks of the Sydney coal basin contains bowlders of rocks which outcrop to the south of the basin and this with the general direction of glacial striae proves that the direction of ice movement in this part of Cape Breton Island was northward. This report is almost entirely economic and gives many details concerning the structure and extent of a number of coal horizons. The New

Ross, Lunenburg County, manganese deposits are described as occurring along a fissure in granite. Calcite and manganese oxide was deposited in this fissure and later movements broke up this vein and formed a frault breccia. Secondary enrichment from surface waters has concentrated the manganese oxide into bodies of workable size and high-grade ore.

3. A few preliminary results are given of investigations of peat bogs near St. John and Moncton, New Brunswick.

Part G. The Platinum Situation in Canada, 1918. By J. J. O'Neill. Pp. 19, map.

The chief platinum-producing areas in Canada are in Ontario, British Columbia, and Yukon. In the nickel-copper ores of Sudbury, Ontario, platinum occurs as sperrylite, the platinum arsenide. In British Columbia platinum is found both in the solid rocks and the gravels. In the solid rocks three distinct types of deposits are recognized—first, in association with chromite in peridotite-pyroxenite rocks; second, in association with chalcopyrite deposits; third, in shear zones in typical granite. In the gravels of Yukon platinum is widely distributed but not in large enough quantities to be profitably exploited for this metal alone.

Canada appears to have possibilities of becoming one of the great producers of platinum. In 1918 only one hundred ounces of platinum were recovered, but probably more than 50,000 ounces of the platinum metals were contained in ores mined in Canada, but not recovered.

J. F. W.

The Silurian Geology and Faunas of Ontario Peninsula, and Manitoulin and Adjacent Islands. By M. Y. Williams. Canadian Geological Survey, Ottawa, Memoir 111, 1919. Pp. 195, appendices III, pls. XXXIV, figs. 6, maps 2.

In this memoir the author gives his conclusions, based on five seasons of detailed field work, on the general Silurian problems of southwestern Ontario. Detailed sections, descriptions, notes on origin and correlation, and complete fossil lists for the various members of the Silurian system are given. Nine diagrams are given to illustrate the conditions of sedimentation during various stages of the Silurian period. Nine new species of brachiopods and one new variety are described. The three appendices contain descriptions of a new species of brachiopod

by Foerste, a new species of crinoid by Springer, and two new species of corals by Chadwick.

The important physiographic feature of the area is the Niagara escarpment which is formed by the outcropping edge of the Niagara dolomite.

The Silurian formations classified on the basis of lithology fall into the following three groups in ascending order: (1) Alternating sandstones, shales, and limestones represented by the Medina-Cataract, Clinton, and Rochester formations and indicating changing conditions of land in respect to the sea. (2) Massive dolomites represented by the Lockport and Guelph formations and suggestive of widespread seas of moderate depth. (3) Saline sediments containing lenses of salt, gypsum, and impure clastic dolomites represented by the Cayugan group which were formed in shallow, practically isolated interior water basins.

The disconformity between the top of the Ordovician represented by the Richmond and Queenston shale, and the base of the Silurian represented by the Whirlpool sandstone is distinct. The Bass Island group of the west and the Akron dolomite of the east are put at the top of the Silurian and the disconformity between these formations and the basal Devonian is also well marked at a number of localities. Breaks in sedimentation occur at the base of the Lockport and Salina.

Chapter vi contains notes on the salt, gypsum, petroleum, natural gas, and other materials of economic importance found in the area.

The report is well illustrated and is a careful, detailed, and concise statement of the Silurian geology of southwestern Ontario.

J. F. W.

The Geography of Maryland. By WILLIAM BULLOCK CLARKE. *The Surface and Underground Water Resources of Maryland, Including Delaware and the District of Columbia.* By WM. BULLOCK CLARKE, E.B. MATTHEWS, and E. W. BERRY. Maryland Geological Survey, Vol. X, 1918. Pp. 553, figs. 96.

Part I is a brief discussion of the geology and physiography, including the Coastal Plain, Piedmont Plateau, and the Appalachian physiographic provinces, climate, flora and fauna, and the natural resources of the state. Among the chief resources may be mentioned coal, clays, building and decorative stones, limestone products, agriculture, and timber. A number of suggestions for physiographic and geologic excursions within the state are included.

Part II is a more detailed discussion of the geology and physiography of the region. The geology is dealt with by physiographic provinces and includes sedimentary, igneous, and metamorphic areas, and, stratigraphically, rocks from pre-Cambrian to Recent. The discussion of the underground water resources, which forms the greater part of the paper, includes an explanation of the general principles involved and local detailed descriptions of the resources by counties. There are appended to the report eleven tables of statistics of various sorts.

A. C. McF.

William Smith, His Maps and Memoirs. By T. SHEPARD, M.Sc., F.G.S. Proceedings of the Yorkshire Geological Society, N. S., Vol. XIX, Part III. Pp. 178.

William Smith was one of the pioneer English geologists in stratigraphic and areal work. The report consists of descriptions of his various maps and writings, the first produced in 1799 and the last in 1827. It includes many reproductions of the original diagrams and charts.

A. C. McF.

Upper Cretaceous of Maryland, Systematic Report. Maryland Geological Survey, 1916. Pp. 1022, pls. 7 (general), 90 (paleontological).

I. The Upper Cretaceous Deposits of Maryland," by W. BULLOCK CLARKE.—Under this heading is included a discussion of the general geology of the Coastal Plain region of the state, to which the Cretaceous deposits are limited, including the physiography, stratigraphy, structure, and conditions of sedimentation. A bibliography and table of distribution of the fauna and flora are also given.

II. "Petrography and Genesis of the Sediments of the Upper Cretaceous of Maryland," by MARCUS I. GOLDMAN. Based upon petrographic and field evidence.—The author finds three types of sediment present, (1) delta type, (2) lagoon type, and (3) open-water glauconitic type. A brief discussion of the origin of glauconite and the methods of petrographic examination is given.

III. "The Upper Cretaceous Floras of the World," by E. W. BERRY. —No attempt at detailed correlations of these widely scattered floras is made. A discussion of the place of origin and subsequent migrations of the great dicotyledonous flora, which makes its sudden and dominating

appearance in the Upper Cretaceous, is given. The flora shows great modernization compared with the Lower Mesozoic horizons. Extensive floral lists are given.

IV. "Correlation of the Upper Cretaceous Formations," by W. Bullock Clarke, E. W. Berry, and Julia A. Gardner.—Complete accordance between the faunal and floral evidence seems to be lacking. The problems involved are discussed in detail.

V. "The Systematic Paleontology of the Upper Cretaceous Deposits of Maryland," by R. S. Bassler, E. W. Berry, W. B. Clarke, Julia A. Gardner, H. A. Pilsbury, and L. W. Stephenson.—Some 325 species and varieties are described of which approximately one-fifth are fossil plants. The majority of these are figured. The volume contains ninety plates of excellent figures. The report is of importance to the stratigraphic and paleontologic world.

A. C. McF.

Recent Publications

—Ross, C. S. Structure and Oil and Gas Resources of the Osage Reservation, Oklahoma. Tps. 20 and 21 N., R. 12 E. [U.S. Geological Survey, Bulletin 686-N. Washington, 1919.]

—Schaller, W. T. Mica in 1918. [U.S. Geological Survey, Mineral Resources of the United States, 1918. Part II: 26. Washington, 1920.]

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—Schofield, S. J. Geology and Ore Deposits of Ainsworth Mining Camp, British Columbia. [Canada Department of Mines. Geological Survey, No. 99, Geological Series. Memoir 117. No. 1773. Ottawa, 1920.]

—Schultz, A. R. A Geologic Reconnaissance, for Phosphate and Coal in Southeastern Idaho and Western Wyoming. [U.S. Geological Survey, Bulletin 680, 1918. Washington, 1919.]

—Schwennesen, A. T. Ground Water in the Animas, Playas, Hachita, and San Luis Basins, New Mexico. With Analyses of Water and Soil, by R. F. Hare. [U.S. Geological Survey, Water-Supply Paper 422. Washington, 1919.]

—Schwennesen, A. T., and Meinzer, O. E. Ground Water in Quincy Valley, Washington. [U.S. Geological Survey, Water-Supply Paper 425-E. Washington, 1919.]

—Scientia, Vol. XXVII, N. XCVI-4 Series II. [Bologna: Nicola Zanichelli, 1920.]

—Smith, George Otis. Economic Limits to Domestic Independence in Minerals. [U.S. Geological Survey, Mineral Resources of the United States, 1917. Part I: A. Washington, 1919.]

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DIASTROPHISM AND THE FORMATIVE PROCESSES

XV. THE SELF-COMPRESSION OF THE EARTH AS A PROBLEM OF ENERGY

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The University of Chicago

The discoveries of the last three decades have led to new views of the constitution of matter, new evaluations of cosmic energy, new estimates of evolutionary rates, and new concepts of the time factor generally. Nearly all the fundamental concepts of geology need some degree of revision in the light of these radical advances. Among the rest there is need to rectify the concept of the earth's compression.

THE CONCEPT OF COMPRESSION IN THE LIGHT OF NEW CONCEPTS OF MATTER

So long as matter was supposed to be formed of minute irreducible atoms, it was logical to assume that when these atoms were pressed into contact there was an end of compression. It was also quite natural to build upon this mechanical concept a merely mechanical notion of the process of compression. The new discoveries, however, lead to the view that the atom is a highly dynamic organization, a complex revolutional system, carrying within itself prodigious stores of energy and a structure as open as a planetary system. The materialistic factors—if

indeed they are really materialistic at all—recede to minute points, and do little more than play the part of carriers of electric charges. The prodigious energies of the atoms seem to be stored in the extremely rapid revolutions of these charged integers and in the fields of force and the polarities which arise from them. The atom itself and all the combinations into which it enters are therefore to be regarded as theoretically compressible to an undefined degree. The old assigned limit vanishes, and no new one takes its place. For aught that is now known, even the nuclei, or protons, and the electrons may themselves be composite dynamical organizations and subject to compression. The fact that a nucleus has a mass 1,800 times that of an electron suggests that perhaps analysis has yet one or more steps at least to take. Compression is therefore to be pictured as the struggle of one phase, or one set of phases, of energy against another phase, or set of phases, of energy, both sets being embodied in motion.

While perhaps it cannot yet be said to be strictly proved that the positive and negative charges are in revolution about one another, there seems to be no other way in which the prodigious energies associated with them can be stored without giving such evidences of themselves as characterize the non-revolutionary activities, distinctions to be considered later. Moreover, the notable successes of the revolutional hypotheses in accounting for observed phenomena leave little room for doubt that they are substantially true, and may be taken as a fairly safe working-basis.

In addition to the evidences of the atoms themselves, the analogies of the larger units of the cosmos lend support to the view that the atoms are revolutionary organizations.

THE CONTRASTED MANIFESTATIONS OF ENERGY

In the great stellar field, where the largeness of things makes visualization easier than in the hidden ultra-microscopic world within, energy manifests itself in two rather distinct kinds of activity. The one is continuous motion in cyclic orbits or spiraloid revolutions running on indefinitely without loss of energy. It is, therefore, conservative and singularly undemonstrative. In the other, the motion is habitually interrupted by reversals and so is

discontinuous and disjunctive, giving rise to diversions and scatterings of energy in oscillatory radiations. This vibratory phase of activity is at once dissipative, agitative, and demonstrative. It has a general destructive tendency, while cyclic motion has a general constructive tendency. However, by weakening old structures, vibratory action prepares the way for new construction. The two types are therefore co-operative as well as antagonistic. The vibratory type has its chief manifestation in the heat-light-X-ray series; the cyclic type, in the planetary-stellar revolutional systems, and in atomic, molecular, and crystalline organizations. In application to material substances, the revolutional type is predominant in atoms, molecules, crystals, and true solids generally; the vibratory type is most manifest in the true fluidal states; in a special sense it may be said to dominate gases.

THE RELATIVE ENERGY-VALUES OF THE TWO TYPES

In the ultimate analysis of all the cosmic states taken together, the revolutional type greatly preponderates in energy-value. This is not in accord with our sense-impressions. It is a rather singular fact that the values of these contrasted phases of energy are inversely proportional to their *obtrusiveness*. Neither rotations nor revolutions are notably demonstrative, while potential energy of position is only visualized by a mental effort, if visualized at all. The rotation of the earth involves a motion of a fraction of a mile per second; its revolution involves a mean motion of 18 miles per second, while its potential energy of position has a value of 356 miles per second. In this only relations to the sun are included; relations to the rest of the cosmos, in which further great, but only partly known, stores of energy are involved, are neglected.

Over against these great but unobtrusive forms of the earth's energy, stand the very impressive vibratory energies of the heat-light-X-ray series, the specially obtrusive and spectacular energies of the cosmos. While the precise sum total of these cannot be given for lack of adequate data, an excessive estimate may easily be made, and this will serve as a limiting value. According to Lane's law the highest temperature of a condensing body occurs at the stage when it is passing from the gaseous to the liquid state.

Let this stage be assigned the earth in its early history to give it a maximum value of the agitative type of energy. It must then of course have extended far outside its present solid surface. The parabolic velocity—the velocity that carries to infinity—at the present surface, is 6.95 miles per second. It is obvious therefore that the mean velocity of the molecules of the earth-substances could not have been so high as this without dissipating the earth. The maximum mean velocity of the earth-molecules must, therefore, always have been appreciably lower than 7 miles per second. We have, therefore, as the respective mean velocities, something less than 7 miles per second for the vibratory energy, something more than 18 miles per second for the revolutional velocity and—neglecting the rotational velocity altogether—356 miles per second for the potential energy. As the mass is the same in all cases, the energy-values are as the squares of these figures. Reduced and combined, the ratio of the vibratory energy of the earth, on the most generous allowance, cannot be more than 1/2600 of that of the revolutional energy, even when a large factor is neglected. The purpose of this comparison is to show the exaggerated importance that has been given to the agitative phases of energy, as also to the gaseous state, in the study of the earth's energy-values. In this, however, we have only considered the megascopic motions of the earth. We have yet to consider the ultra-microscopic phases in which prodigious energies are even more unobtrusively concealed.

To approach the ratio between the dissipative and the constructive classes of energy in the earth-matter itself, let the familiar case of a bowlder on the surface be taken. Let it have the mean temperature of the earth's surface, say 15° C. Its absolute temperature will then be about 288 centigrade degrees. This represents a linear extension of about .0057. All the rest of its extension represents the work of constructional energy—here interpreted as revolutional energy—except the space occupied by the atomic nuclei and the revolving electrical charges. While the total value of the energy of the revolving constituents of the atoms is undeterminable at present, it is certain that it is almost incomparably greater than that of the 288° C. temperature.

When this atomic energy, which is even more unobtrusive than the energies of celestial revolution, is added to the macroscopic energies, the disparity mounts up to a very high figure. The agitative energies that so deeply impress our senses are really little more than trivial, relatively, in the true cosmic scale.

Now, the resistance that is offered to the compression of an earth made up of solid matter springs mainly from the forces that determine the constitution of this matter. The analysis of these constitutional energies, as now interpreted, involves the electronic revolutions, together with the fields of force and the polarities that spring from them. These may not be all the forces involved—very likely they are not all—but they form the truest picture now available and they may be taken as representative. They are herein made a working-basis, subject to correction as additional light is disclosed.

THE RELATIVE ENERGY-VALUES OF THE POSTULATED EARTH-FORMING NEBULAE

In estimating the potential energy of the nebular matter which, by hypothesis, was condensed to form the earth, in each of the two representative views, the planetesimal and the gaseous or quasi-gaseous, it is assumed, in both cases, that the earth was formed in essentially its present position and relations in the solar system.

In Article XIII of this series,[1] a conservative estimate of the belt occupied by the planetesimals that were later to form the earth, gave it a space-value of 9×10^{23} cubic miles. The gaseous nebula that was to form the earth, measured at the time it first came into self-control and was most extended, had a volume less than 3.5×10^{18} cubic miles. The ratio is about 250,000 to 1. The vastly superior space occupied by the planetesimals, however, does not carry proportional value in potential energy. Its importance chiefly lies in the mode of support of the planetesimals and in their modes and rates of assemblage.

CONTRASTED MODES AND RATES OF ASSEMBLAGE

The modes of concentration were radically different. The planetesimals were sustained in their orbits by velocities of a

[1] *Jour. Geol.*, Vol. XXVIII (1920), p. 678.

mean value of about 18 miles per second. Thus sustained, they only joined the collecting nucleus as variations in their orbits brought them into conjunction with it—a slow process, occupying perhaps two or three billion years.[1] The intervals between the infalls of the planetesimals were, therefore, such that nearly all the heat of their impact with the atmosphere and with the earth's surface was lost before they were buried by added material. The growth of the earth was thus made by the slow accumulation of essentially cold, solid particles mixed at random.

On the other hand, if the earth-forming nebula be assumed to have been gaseous and to have descended along the gaseous line, its volume was sustained by collisions and rebounds of the constituent molecules, and it contracted as fast as the loss of this interaction, i.e., the loss of heat, permitted. Under Lane's law the maximum temperature was reached at the stage when the gaseous body passed into the liquid state. As radiation follows the law of the fourth power, the collapse was relatively rapid; at the most it cannot be assigned more than a few million years.

Enormous losses of energy would be suffered in either the planetesimal or the gaseous mode of assemblage, and so we must take up the question of the earth's primitive energy presently from the opposite point of view: What energy-values were *left* for the evolution after the earth was able to make a record of its own compression? The point of most importance here is the radical difference in the respective factors that controlled the self-compression which followed the nebular concentration. It is obvious that the gaseous descent was controlled by *heat* and that this remained the master factor in the shrinkage of the earth after it became a white-hot molten globe. In the self-compression of the earth built of solid planetesimals, or planetesimal dust, *solidity* was the primary resisting-factor that held the compression in check. The energy-factors in this case were those to which the solidity was due. These are herein interpreted as revolutional phases of energy together with their derivatives. Heat in one case and solidity in the other were then the master factors in the

[1] See "The Rates of Planetesimal Infall," Article XIII, *Jour. Geol.*, Vol. XXVIII (1920), pp. 677 ff.

compression process. In the latter case, the heat generated in the course of the compression was secondary to the revolutional energies. The special courses taken by both the primary and secondary energies become therefore vital elements in the compressional process; to these we shall presently turn.

As indicated above, to form estimates of the energy-values that were inherited by the earth at the stages when it began to make its automatic record of self-compression, it is necessary to enter into a more specific analysis of its status in the two hypothetical cases.

COMPARATIVE VALUES OF ENERGY AVAILABLE FOR DIASTROPHISM

The deformations of the earth are the most available test of the energies that entered into its self-compression, though by no means the only test. There is now no ground to doubt that the diastrophism was large, whatever estimate may be made of its precise value. There must have been enough energy in an available form to actuate the distortions involved, and this energy must have been properly distributed in time and place. The sources of this energy need therefore to be considered in respect to their availability, as well as their adequacy. Fortunately, the problem for all cosmological lines of descent seems to center in the alternative: Was the earth assembled in a fluidal condition dominated by heat, or was it built up gradually by accessions of small fragmental matter in a cool, solid, highly mixed state? If there are tenable hypotheses of an intermediate sort, the considerations that apply to these type-views can easily be adapted to them.

The chief energies available for the evolution from this point on, are (1) the residue of the potential energy of position, except of course what still remains potential; (2) chemical and physical combinations, readjustments, and reorganizations, so far as conditions permitted them to take place during the compression; and (3) the disintegration of radioactive substances, including any other changes in atomic constitution that may have taken place, if any. These atomic factors may possibly have some relation to the extreme stresses that arose from compressional action, but as

there is now no evidence of this, they must be treated under a head of their own.

1. *The period of compression.*—If the earth remained fluidal until all its rock-substance was condensed into a globe, none of the energy lost in the assembling was available for making the observed diastrophic record, since this could only begin after consolidation began. If, on the other hand, the earth was built up of small solid accessions loosely laid down, these must have begun to suffer compression and distortion as soon as one layer was laid on another. The distortional process must in this case have run on thence through the whole history of growth. The compressional and distortional actions were furthermore brought on very gradually and great lapses of time were available to meet the growing stresses by the resources of readjustment, reorganization, metamorphism, and diastrophism.

2. *Availability of the main compression.*—If the earth was assembled in a fluid state, the interior underwent the full measure of fluidal compression from gravitative action before it could make any diastrophic record; little more than the effects of cooling remained available for deformative work after solidification took place. If, on the other hand, the material of the earth was added slowly in a loose, solid state, the main compressive effects entered into the record; for while the distortions in the deep interior would never be accessible, they must have been at all stages the foundation on which the later accessions were built and hence they gave direction to, as well as participated in, the stress effects that arose at every subsequent stage in the increase of mass. They must still continue to participate in the effects of all the more general changes in gravity.

3. *Chemico-physical combinations, readjustments, and reorganizations.*—If the earth remained fluid and convective until fully assembled, almost ideal opportunities for chemical combination and physical adjustment, as well as chemico-physical reorganization, would have been offered, except in so far as the heat itself may have restrained such action. To this extent the chemico-physical resources should have been exhausted before they became available for diastrophism. But if the earth were built up of solid particles

of various sorts mixed by the chance of infall, it would offer almost ideal conditions for recombination, readjustment, and reorganization, which, in this case, would run hand in hand with diastrophism and contribute to it.

4. *Relative exhaustion of potential energy by segregation.*—If the earth was fluid until fully assembled, there should have been facilities for the arrangement of the earth substances in concentric layers according to specific gravity. This would have been a special means of reducing to the lowest terms the potential energy that might otherwise have remained available for deformative work after solidification made a diastrophic record possible. If, on the other hand, the matter remained a heterogeneous mixture so far as intrinsic heaviness was concerned, a corresponding amount of potential energy remained available for the diastrophic record. In so far as segregation by gravity took place during the compression of the mixed solid mass, it co-operated with other deformative processes and left its effects in the record.

5. *Relative exclusion or retention of gaseous constituents.*—If the earth remained fluid and convective until fully assembled, the gaseous constituents should have had favorable opportunities for escape and should have been impelled to escape by the very high heat, so that only such quantities as were required to balance the partial pressures of the same constituents in the atmosphere should have remained to take part in vulcanism later.[1] On the other hand, if the earth was built up by solid particles added slowly to the surface and subjected to weathering and to mixture with air and water, as it was gradually buried, the complex should have afforded almost ideal conditions for the evolution of volcanic gases when it was later subjected to heat and pressure. The phenomena of the moon are especially instructive in this respect, for the gravity of the moon is insufficient to hold free volcanic gases even in its present cold state; much less then in a hypothetical molten state. No equilibrium factor should have been retained in this case. But the evidences of vigorous explosive action on the moon are very pronounced.

[1] Rollin T. Chamberlin, "The Gases in Rocks," *Jour. Geol.*, Vol. XVII (1909), pp. 565–68.

6. *The distribution of the radioactive substances.*—If the earth were assembled in a fluid state, the radioactive substances should have settled toward the center because of their high specific gravity, or else, if convection prevented this, they should have been distributed sub-equally through the whole mass. There should at least have been no concentration of such heavy material in the upper layer. But the special investigators of the subject agree that if the whole earth were as rich in radioactive substances as its accessible portion is, the heat generated would be many times greater than the heat now conducted to the surface and radiated away. Were this true, the earth should have been growing hotter all through its history and no shrinkage at all could be assigned to cooling. On the other hand, if the earth were built up of heterogeneous clastic matter that carried its chance portion of radioactive particles, and if these, by their heating action, liquefied the most susceptible matter immediately enclosing them, and if such liquid matter were then squeezed to or toward the surface by the powerful extrusive agencies that belong to a solid earth, the radioactive substances would be concentrated in the zone of lodgment of these igneous portions. This limits the radioactivity to a degree that seems to fit the observed facts and the theoretical intimations of the case. It is quite obvious that, so far as deformative effects assignable to cooling are concerned, the hypothesis of a molten earth is seriously embarrassed by this newly discovered source of heat superposed on an already embarrassing inheritance of heat from its earlier history, while under the hypothesis of a cold-grown solid earth, it is a welcome agency.

The combined import of all the preceding considerations leaves the fluid earth embarrassingly short, if not fatally short, of resources of energy available for making the observed diastrophic record, while the planetesimal earth is much more amply, and apparently quite adequately, supplied with such energy, and this becomes available in such a slow way as to give great allowances of time for the increments of compressive stress to work out their adjustments and easements along metamorphic and diastrophic lines.

THE INTERCHANGES BETWEEN THE TWO BASAL TYPES OF ENERGY

Before taking up the special modes of the compressional process, the interchanges between the two basal types of energy need consideration. The constructional and the agitative phases of energy are not only interchangeable but interchanges are persistently taking place on the surface and within the earth, and these interchanges play a vital part in the process of the earth's self-compression. The proper recognition of these is indispensable.

Exchanges between thermal and mechanical energy are too familiar to need notice; they are a basal feature in modern industry. But exchanges between agitative and organizing energy, i.e., between vibratory and revolutional energy, as such, though they may not differ in essence from the well-recognized interchanges, need a word of emphasis. Some of these changes from the agitative to the constructive are even more familiar than the mechanical changes, but interpretation has not given them the value to which they are entitled. We know that the grass and the trees grow, but we easily overlook the fact that such growth is a widespread and important endothermal process. It belongs to the unobtrusive class and does not enforce attention. The prairie fire and the holocaust of the forest, the complementary exothermal process, command our lively attention. The unobtrusiveness of endothermal action is likely to deceive us as to the balance between interchanges of energy in nature. The problem in any special case is to determine the balance between opposing actions. There is, however, no doubt as to a real preponderance of exothermic action on the earth's surface. When lavas come up from below, they usually undergo exothermic reactions to a greater extent than endothermic reactions. This in itself raises the question whether endothermic reactions are not preponderant in the region whence the lavas come. Van Hise,[1] Leith,[2] and their associates, have shown by the extensive collection and study of data from the full

[1] C. R. Van Hise, "A Treatise on Metamorphism," Monogr. XLVII, *U.S. Geol. Surv.* (1904).

[2] C. K. Leith and W. J. Mead, *Metamorphic Geology*, Henry Holt and Co. (1915). See particularly the chapters on "Katamorphism" and "Anamorphism" in both works.

range of the accessible terranes, that while exothermic reactions preponderate in the outer or katamorphic zone, the preponderance is reversed below and endothermic reactions take precedence in the anamorphic zone. It is to be noted that while katamorphic action, exothermic action, and the lowering of density, commonly go together, as also anamorphic action, endothermic action, and rise of density, they do not invariably coincide; and further, that none of these necessarily excludes the others from any horizon. The essential question is not one of exclusive action but of preponderant action.[1] It is in the natural order of things that in the great contact zones between the atmosphere, the hydrosphere, and the lithosphere, there should be a trend of energy toward its agitative phases, and that in the stabler solid zones below there should be a compensating trend toward the constructional and the persistent, without limiting either zone to one type of action. In terms of the two basal types of energy, exothermic action, on the average, involves a change of organizing or revolutional energy into vibratory-dissipative energy; while endothermic action is commonly the reverse.

THE CONDITIONS THAT DETERMINE THE INTERCHANGES

In a very broad sense, open conditions and freedom from pressure or other forms of restraint, favor exothermal reactions, while confinement and pressure favor endothermal reactions. Any form of crowding, even self-stress, naturally tends toward *divergence* of energy into the various paths available to it, since this affords relief. The higher the stress, the more it forces itself into unoccupied paths. Concentrative stress, therefore, favors the passage of a portion of the energy along endothermic lines and the formation of dense substances; while dispersive stress, low stress, and no stress, are less compulsory and give exothermic action freer scope. Apparently crowding is not confined to imposed stresses, but arises from what may be styled the self-stress of the activity. A small mass of gas in open space exerts little interior stress upon

[1] Compare C. K. Leith, "The Structural Failure of the Lithosphere," Vice-Presidential Address, Geol. Soc. Amer., *Science*, N.S., Vol. LIII (March 6, 1921), pp. 205–7.

itself, and only the larger vibrations are in evidence, but if the mass grows indefinitely the internal self-stresses increase and there appear in succession the shorter and more intense vibrations ranging up through the whole gamut of vibrations to the X-rays and doubtless beyond. There is, in this, increased pressure, of course, but the activity itself is increasingly *divergent* as well as increased in amount. However this may be interpreted, there is a growing complexity of vibration, and it seems to be a safe generalization that growing mass and growing internal pressure are attended by increase in the diversity of phases assumed by the compressional energy; in other words, there are more varied partitionings of the energy and it takes a larger number of paths, including more frequent interchanges between the endothermic and exothermic phases. As there is thus crowding in various directions for easement, the direction that gives greatest relief from the stress imposed by the environment naturally becomes a predominant trend. Where there is high pressure and it is unescapable, the line of relief is the passage of energy into a constructional form that gives additional density. Where the pressure is weak or absent, an expansional or dispersive form of energy may be more efficient in giving relief. Both forms are likely to be present and to co-operate with one another in any pronounced case.

THE TESTIMONY OF PRESENT INTERNAL STATES AS TO THE DOMINANT DIRECTIONS TAKEN BY ENERGY IN THE INTERIOR

Tidal[1] and nutational[2] evidences concur in indicating a higher degree of rigidity and elasticity in the interior, taken as a whole, than in the outer shell. Seismic waves add very specific confirmatory evidence, so far as the outer seven-eighths of the volume of the earth is concerned. The seismic evidence for the remaining central part is as yet obscure, and is differently interpreted by the special students of the subject. In a general way, the whole of the interior is covered by the tidal and nutational evidences. These favor the interpretation of the central part as highly rigid and elastic, since

[1] A. A. Michelson and Henry G. Gale, "The Rigidity of the Earth," *Jour. Geol.*, Vol. XXVII (1919), pp. 585–601.

[2] W. Schweydar, "Die Elasticität der Erde," *Naturwissenschaften*, Part 38. Potsdam, Germany (1917).

these qualities fit the general import of the evidence, but for the present it is prudent to leave the question of the state of the center to be settled in the future. It is to be observed that the increasing density of the interior tends to dampen the speed of the seismic waves, and that correction for this effect must be made in deducing the inward increase of rigidity and elasticity from the seismic records. When allowance is made for this, the generalization that rigidity and elasticity are notably higher in the interior than in the outer shell is put beyond serious question. This means that in the partition of the compressional energy between those phases that increase the rigid elastic attachments of the molecules to one another and those phases that weaken or destroy these attachments, the former have been favored in a marked degree. This is testimony of a most cogent sort. By interpretation, this signifies that only a minor part of the compressional energy took the vibratory form in the interior, the major part taking the revolutional or constructive form, and that in doing this it served to promote compactness and the strength of hold of the constituents on one another.

DID THE ORGANIZING ENERGY EFFECT CHANGES IN THE CONSTITUTION OF THE ATOMS?

Lest the seeming needs of the case bias us toward one conclusion rather than another, let us hasten to note that the mean density of the earth, compared with the probable density of the original matter, is such as to offer no real ground for bias in favor of atomic construction, for the higher density of the interior is fully accounted for by the density gradients that arise from metamorphism in the zone of observation. Atomic construction, if invoked as an aid, might as easily render the interior mass too dense, as to help explain the density as it is. Nor is there more than uncertain evidence bearing on the atomic question; but the matter is too important to be ignored in a discussion of the effects of compressional energy.

The most remarkable of known exothermal effects connected with rock-substance springs from the spontaneous atomic disintegration of the heaviest known elements. No evidence that this

disintegration has anything to do with relief of pressure, such as might be assigned to their rise from the interior, is now available. Any such possibility must be left to the revelations of the future. It is logically necessary, however, for one who believes in the indefinite cyclic persistence of the cosmos, to suppose that the present exothermic action is the reversal of an endothermic process that gave these elements the stores of energy they are now so persistently and systematically discharging. The place and time and conditions of this storing action are altogether open questions. By interpretation, the energy now being given out springs from intense revolutional action, for revolutional motion seems to be the only probable way in which such prodigious energies can be stored in so unobtrusive a state and given out so regularly and systematically and in such concentrated forms. The storing process must probably have involved somewhat similar forms and intensities of action. One of the most common speculations as to the place and conditions of this storage process, locates it in some center of great stress where pressure and heat co-operated. This should perhaps be amended by recognizing that the more intense vibratory agencies of the X-ray end of the series were even more probable agencies, because their motions were more nearly commensurate with the minute and swift revolutions that are supposed to store the energy in question. The center of the earth is possibly a place of the right type, but it belongs to an inferior order compared with the centers of stars, unless solidity counts for something. In this case the center of the earth might have a preferred place, since our planet is among the largest of known solid bodies. An alternative speculation places the origin of the radioactive substances in the outlying regions of space.

There is perhaps a suggestion of general atomic change in the remarkable phenomena of thermionic emission, contact potentials, and photoelectric action. These seem to imply that there is some kind of commensurability between the extremely intense oscillations of the vibratory activities and the orbital periods of the electrons, so that effective interaction and perhaps interchange takes place between them. Commensurability is perhaps the property by which interchange is effected between the minutely

vibratory and the minutely revolutional phases of energy. These speculative suggestions have little value beyond helping to make it clear that it is by no means safe to assume that atomic construction and destruction are not common functions of the interior.

WHAT AMOUNT OF COMPRESSION IS IMPLIED BY THE MEAN DENSITY OF THE EARTH?

As already noted, there is no need to push appeal to the organizing functions of energy in the interior so far as to assume the building up of atoms for the sake of explaining the higher density of the earth's interior; indeed, if there is any constructive work of that sort, the increase above the decrease of density cannot go very far without making the mean density too great to fit the evidence. If we assume that the primitive matter had the meteoric density of 3.69 adopted by Farrington, and compare this with 5.53, the mean density of the earth adopted by Moulton, the mean increase in density due to compacting, reorganization, atomic change, etc., is only about 50 per cent. Or if, to assume an improbably low figure for the density of the earth's original matter, we take the moon's mean density, 3.34—assuming that the effects of compression at the moon's center are offset by the porosity of its outer part—the increase in the earth's mean density would be only a little over 65 per cent. In either case, or on any plausible assumption, some part of the compacting must be assigned to mechanical compression, so that the increase of density assignable to reorganization under the special conditions of the interior is not very large.

THE INTIMATIONS OF THE DENSITY GRADIENT IN THE ZONE OF OBSERVATION

Geologists have been at great labor to compile thermal data from mines and deep borings that they might deduce from these a temperature gradient that would throw light on interior conditions, but the same line of attack on the rising density of the interior seems to have been overlooked. It is to be recognized, of course, that neither of these gradients can be projected to the center of the earth without reservation, for both curves probably fall off notably in the interior, but the density curve is probably as trustworthy a

guide as the temperature curve, for, in the planetesimal earth, both arise from compression and its direct and indirect consequences.

After an elaborate study of the most reliable data, Dr. H. S. Washington thus sums up his conclusions in a recent paper:[1] "I am inclined to place the average density of the crust at about 2.75 at least for the uppermost shell, while that of 2.80 would probably be nearer the truth for an average of any considerable depth, say 20 miles or more." The mean depths of these two shells can scarcely be more than 8 or 10 miles apart. The rise of density in this little difference of depth, if projected to the earth's center, would give a density there nearly twice that computed from the classic law of Laplace, or from the law of Roche specially formulated to meet the astronomical requirements. No account is here taken of mechanical compression, for the specific gravities adopted as the basis of the estimates were all taken under atmospheric pressure. Much less was any account taken of hypothetical quantities of metals or other specially heavy material, for both these shells are formed of common rock.

The two elements most common in the zone of observation, oxygen and silicon, often unite to form tridimite in the outermost shell but not in the plutonic rocks, where the same elements appear as quartz. This distribution is commonly assigned to differences in the physical conditions of the two horizons, especially differences in pressure. The specific gravity of tridimite ranges from 2.28 to 2.33, while that of quartz is 2.65. There is thus a rise of density of 15 per cent, so far as these minerals are concerned, between two horizons both of which lie in the limited zone made accessible by deformation and denudation.

The most instructive and suggestive data, however, are found in the progressive stages of increase in density developed in several different kinds of silts as they pass into various kinds of schists, and thence, in part, into a group of heavy minerals of which the garnets may be taken as types. The compression of the silts into shale may be neglected since a notable part of the increased density

[1] H. S. Washington, "The Chemistry of the Earth's Crust," *Jour. Franklin Inst.*, December, 1920, p. 804.

of the shale was due to the mechanical elimination of porosity. In forming the schists there was true reorganization with increased density, and still later there was further partial reorganization into much heavier minerals of the garnetic group. In the case of the garnets there is a rise in density from the schist minerals which formed them of 36 to 84 per cent, as pointed out by Van Hise.[1] In these cases the rise of density is unequivocally the result of the metamorphic reorganization of very common and representative kinds of material. It is to be noted further that one reorganization follows another even in this limited zone.

These specific cases show the possibilities and the actual tendencies to increase of density by metamorphism, quite apart from mechanical compacting. They point to the very significant fact that the chief density effects are to be sought in metamorphism rather than mere mechanical compacting by pressure. The crux of the compressional problem therefore lies in metamorphic reorganization, in selective liquefaction, and in the extrusion of magmas. The concrete task thus imposed is the tracing of the paths followed by the compressional energy and the study of the kind of work each phase of energy does. The specific phenomena to be explained are (1) the rising density, rigidity, and elasticity of the interior material; (2) such a distribution of density as to satisfy the intimations of the precession of the equinoxes and the nutation of the poles; (3) the amounts and kinds of diastrophism recorded in the structure of the earth; and (4) the protrusion and persistent maintenance of the continents and the complementary depression of the ocean basins, as well as the special configurations of the earth. It is not sufficient to explain these separately by isolated postulates. unless these are shown to be mutually compatible and connected with a common origin; these features are to be explained as a co-ordinate group of phenomena arising from a common origin and a common line of dynamic procedure.

THE SPECIFIC PATHS OF THE COMPRESSIONAL ENERGY

In the following analysis, it is to be understood that the earth is assumed to be, and to have been at all stages after the formation

[1] C. R. Van Hise, *op. cit.*, pp. 205 ff.

of its collective nucleus, a solid body, built of heterogeneous planetesimal matter at the start, that it was subjected to a slowly growing gravitative pressure whose total accession was spread over a period of the order of two or three billion years, and that there were large allowances of time for metamorphism and for the adjustment of the strain arising from each increment of pressure. On these assumptions the chief partitions and paths of the compressional energy, seem logically to have been as follows:

The first step was the partition of the initial increment of energy by the passage of a part of the stress into strain, while another part took the thermal form. So long as the strain lasted, its energy was stored or latent. A varying but large measure of energy seems thus to have been stored all through the geologic ages. It appears from stratigraphic evidence that the strain-limit in the sub-surficial material has been high enough at several periods to permit the accumulation of stored energy sufficient to actuate declared deformative "revolutions" in spite of such partial easement as may have taken place, during the stages of accumulation, from the milder forms of idiomolecular action about to be described.

The second step in the compressive process was the co-operative action of the stored energy of strain and the agitative thermal energy. The latter aided change by loosening the fixed elastico-rigid attachments of the molecules, the essential properties that gave rigid symmetry to the crystals and solidity to the amorphous fragments. The hold of crystals and of clastic fragments upon their constituent molecules was unequal because some of them lay at the angles, or on the edges, or on sharp curves of the little masses, where fewer other molecules supported them. So also the strains arising from pressure upon the interlocking crystals or fragments was greatest on these outlying molecules. The particular molecules thus least securely held and those most severely strained, yielded first. This eased the strain on these particular points and threw the stress on new molecules; from this, new action of like order arose and the process was essentially repeated.[1]

[1] In many cases, especially near the surface, solution and chemical reaction greatly aided molecular transfer and crystalline reorganization, but these are accessory agencies rather than factors of the compressional process, as such.

The detached molecules of this first action, responding to such stress as was then felt by them in their relatively free state, took the lines of least resistance until they reached some point where the organizing force of some crystal, so situated as to be able to grow, brought them under control and reattached them with due orientation. Such reattachments were obviously conditioned by the balance of strength between the crystalline force and the weakest phase of the general pressure. As a result the growing crystal extended itself most in the line of least resistance and co-operated with other crystals of like situation in developing parallelism of structure.

It is obvious that such individual actions on the part of single molecules acting by themselves, and acted upon by special stresses, could take place while as yet the *general* strain was far below the strain limit and *general* detachment could not take place. If the pressure came on slowly enough, the whole crystal or fragment might be broken down in this piecemeal way while the general strain was below the mean strain-limit. As only a few molecules were in transit at any given time, the mass as a whole would remain solid throughout the process.

The action was thus the special work of individual molecules, each suffering its own strain and playing its own part in its own way, i.e., it was *idiomolecular*. The process is sharply distinguishable from the common movement of all molecules, such as usually takes place when liquids flow. The process may be studied to advantage in the granulation of snow at temperatures that inhibit liquefaction.[1]

So long as the stress and strain were mild, the foregoing action was obviously slow and had rather narrow limitations. But with notable increase of stress, giving rise to increase of strain, and increase of heat, the process appears to have been hastened and given a tendency to collective action in parallel lines, planes, or belts, doubtless because resistance was less effective against such

[1] C. S. Peet and E. C. Perisho, working with the writer in the winter of 1894, found by daily micrometric measurements of many granules, that the larger ones grew every day whether the temperature was at, above, or below 0° C., the growth apparently taking place at the expense of the smaller, more sharply curved granules.

united action. This apparently went so far in some cases as to verge toward general simultaneous molecular action, but analysis seems to show that it remained idiomolecular in actual method. From such quasi-collective but really idiomolecular action, cleavage, schistosity, and other forms of structural parallelism arose. In glaciers this idiomolecular type of action seems to range from snow granulation to the point where fracture takes place and the movement becomes a massive shear.[1]

By further increases of pressure, the strain limit *along selected lines* was reached and definite fracture and shear took place, or else the whole mass was crushed to fragments. In either of these cases, the action became diastrophic rather than metamorphic.

It has been very generally held that when such depths and pressures are reached as to inhibit fracture, general movement of the molecules over one another after the manner of liquids must take place. The original idea of "rock flow" seems to have sprung from this notion. In a highly rigid body, such a general movement of molecules upon one another requires the breakage of all the bonds between molecule and molecule and so *involves a maximum of force.* Moreover, this force must come from *differential* stress, since *balanced* stress, within limits, forces the molecules into closer and stronger relations. The supposed "flow" seems improbable except when true liquidity, which destroys the special rigid bonds of the solid state, is brought about. So far as differential stresses affect the solid matter of the interior, easement by idiomorphic action, either collective or isolated, seems to require much less energy and is hence more probable. With the open structure now assigned matter, and with the abundant evidence that molecules really collect into crystals in the midst of rock that seems quite solid, and with other phenomena giving evidence that in some way molecules creep through solid matter, there seems no substantial ground for excluding idiomolecular action from the deep interior.

[1] It was from the study of the granulation of snow, the growth of glacial granules, and the development of schistosity in the glaciers of Greenland, in 1894, that the method and importance of this individual action of molecules, while the mass remained solid, was first realized by the writer. "Glacial Studies in Greenland," Presidential Address, *Geol. Soc. Amer.*, Vol. VI (1895), pp. 209–14.

In tenacious solids where impact takes place at high velocities and with prodigious force, as in the case of a steel target struck by a solid shell, there is no time for idiomolecular action, and very little for any form of selective or metamorphic action, and so all molecules are apparently caused to move over one another in a way that is scarcely distinguishable, if at all, from real flow. The slowness of the increases of stress in the interior of the earth, however, is thought to put deep-seated diastrophism in a quite different category from this velocity-stress action.

Although fluidal action is placed in a secondary order in the evolution of a planetesimal earth, the formation and extrusion of magmas play a very important function in its compression, but that must be left for a later article.

FIELD OBSERVATIONS IN NORTHERN NORWAY BEARING ON MAGMATIC DIFFERENTIATION

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Recent years have witnessed a marked development in the understanding of differentiation processes. Petrographers have tried to get away from the purely theoretical considerations about the matter and to harmonize the conclusions from the field observations with the results of synthetical experiments on silicate minerals and their crystallization obtained by the Geophysical Institute at Washington and others. Although these experiments are still far from covering all subjects involved, and although the multitude of field observations from most parts of the earth are often contradictory, there seem to be certain lines of development which prove promising.

The results now generally converge toward the conclusion that an ordinary fluid silicate magma, without concentrated mineralizers, is not capable of splitting up into two magmas mutually insoluble or with limited solubility. Accordingly, the main part of the differentiation processes is transferred to the period of crystallization, resulting in considerable restriction of possibilities.

According to this supposition, naturally the first mode of separation to be considered was that of the heavier crystals from the lighter ones and from the still fluid magma by gravitative settlement. In a number of instances this sort of differentiation has definitely been proved to occur. But it also became obvious that this sort of separation could not alone account for a great many differentiation processes actually observed.

The newer theory of squeezing differentiation—quite as well in accord with the latest results—seems to be capable of a more general application in those very frequent instances where lateral

[1] State mining geologist, Geological Survey of Norway.

pressure prevailed during crystallization. This theory was suggested by A. Harker, and especially after it had been formulated and developed by N. L. Bowen[1] it has aroused keen interest among petrographers. In the present paper I have attempted to give some examples from northern Norway bearing on this sort of differentiation.

SHORT REVIEW OF THE GENERAL GEOLOGY OF CALEDONIAN

As known, the Caledonian mountain chain—of late Silurian to early Devonian age—traverses the whole length of Norway from SSW. to NNE., in its northern part occupying nearly the whole breadth of the country. The axis of the chain forms a marked geosynclinal depression of the old Archaean Scandinavian shield and of the pre-Cambrian peneplain. The depression is filled with a very thick series of sediments, strongly folded and metamorphosed.

It is supposed that the whole of Scandinavia has been covered by the Cambro-Silurian sediments, remnants of which are found at many places above the Archaean rocks. They are unmetamorphosed and unfolded wherever well beyond the Caledonian folding region. Toward this old mountain chain, first folding sets in, then we meet an increasing degree of metamorphism, which attains its maximum at the axis of the chain. In the same direction the thickness of this series of sediments increases markedly. In the eastern, unfolded zone (especially in Sweden) the total thickness is generally only some few hundred meters, in the folded but unmetamorphosed Kristiania region the thickness surpasses 1,000 m., and toward the mountain chain it reaches several thousand meters.

Although no determined fossils have yet been found in the metamorphosed sediments of *northern Norway*, there are several reasons to believe that the sediments here represent the same Cambro-Silurian series. The original sediments in the geosyncline were extensive layers of slates, marls, limestones, and dolomites, with subordinate sandstones. The metamorphism has been

[1] N. L. Bowen, "The Later Stages of the Evolution of the Igneous Rocks," *Jour. Geol.*, Vol. XXIII (1915), suppl.; "Crystallization-Differentiation in Igneous Magmas," *ibid.*, Vol. XXVII (1919), pp. 393 ff.

very strong throughout the region, producing garnet-mica schists, marbles, quartzites, etc., but generally not with the development of lime-silicate minerals.

In this region there has not been disclosed any unconformity or discontinuity in the sedimentation.

During the Caledonian folding this series was intruded by great masses of eruptive rocks, especially in the axial part of the chain. Most of these are ordinary granites.

There also occur in considerable quantities femic eruptives, very intimately intruded in the schists, and like these thoroughly metamorphosed to amphibolites and different eruptive gneisses. They are much differentiated, the last products generally being soda-rich granites (or Trondhjemites).

Moreover, we find as isolated fields more extensive areas of gabbroidic eruptives, less metamorphosed, sometimes nearly fresh. They were also intruded during the Caledonian folding period and are chemically nearly associated with the former group, from which they do not differ very much, if at all, in age. They also show marked differentiation and on account of their relative freshness afford very interesting petrographical material.

All the eruptives seem to belong to the same cycle of orogenic intrusions. Original lavas or tuffs, which occur in great masses in the Trondhjem district farther south, have not yet been identified among them.

The intrusions have everywhere occurred under a heavy load of sediments, and the whole complex must have been heated to a considerable temperature.

Outside the real root of the mountain chain all eruptives seem to have been intruded completely parallel to the schists. Especially in the district to be considered here, we nowhere find crossing eruptive dikes or other eruptive bodies cutting the schistosity (excepting the pegmatite dikes). It seems to have been a general rule that the magmas, after intrusion, were subject to *magmatic migration* between heated schists, from the root of the mountain chain outward. In this way the magmas may have wandered for very considerable distances from the places where they broke into the schists, until they came to rest. The moving force was induced

by the orogenic folding itself, and lateral pressure existed throughout the crystallization period.

THE RAANA NORITE FIELD (FIGS. 1 AND 2)

This eruptive body, to be especially considered here, is situated at the south side of the Ofoten Fjord, west of the known harbor of Narvik, at 68°20′ latitude. It has recently been closely investigated by the author on account of discoveries in the last years of extensive but poor deposits of nickeliferous pyrrhotite.

In the very rugged country with steep mountains rising directly from the sea and differences in height of more than 4,000 feet, the whole eruptive mass is exceedingly well exposed, and the results of differentiation can be followed in all details.

The norite is injected into a thick series of garnet-mica schists with some interstratified layers of marble. The injection is parallel to the schistosity and forms a lens-shaped body with a thickness of about 3,500 m. and a diameter of about 12 km. The relative thickness of the lens is greater than is generally the case in this sort of intrusions, and is supposed to be due to the influence of some E.–W. folding axes. The section of the eruptive body with the present surface has an area of 67 km.^{2}, 3 km.^{2} of which has been cut off by erosion. No offsets or crossing dikes occur in the surrounding schists.

The very first investigation teaches that the norite field is not homogeneous throughout, but is composed of a central mass of *quartz-norite* and a very considerable and continuous marginal zone of more femic *normal norite*, occupying the border against the schist all around the field to a width of one-third to one-fourth of the whole diameter.

To understand the reason for this, it is important to know the tectonic position of the eruptive mass. One might of course be tempted to believe that the field after differentiation might have been thrown down in an inverted fold, and that the basic border zone accordingly should represent the lower part of the magma basin, separated out by gravitative differentiation in the same way as is the case in the Sudbury field. This, however, is definitely

proved not to be the case. The norite body forms a lens in its normal position between the schists, with a mean dip of about 30°, and the basic border zone is quite as well developed in the upper as in the lower part of the lens.

Between the quartz-norite and the normal norite there is no eruptive contact, but a gradual though rapid transition.

As a third group we may unite all the olivine-bearing rocks: *lherzolites*, *troctolites*, and *olivine-norites*, generally very rich in olivine and sometimes nearly of dunitic composition. They occur as very numerous, greater or smaller bosses and bands in the normal olivine-free norite of the marginal zone, but never in the central quartz-norite.

In the marginal zone they occur irregularly distributed all through it from the outer contact with the schists toward the inner border against the quartz-norite and all around it, quite as numerous in the upper as in the lower part of the eruptive. They strike one as being swimming bodies in the norite magma. It is easily proved that they are neither younger intrusions nor older inclusions, but are very nearly of the same age as the environing norite.

While the above-named groups of rocks are all very nearly related and form a stepwise but nearly continuous series without definite eruptive contacts, there is chemically and tectonically a gap between these and a last group of eruptive rocks. The latter form well-defined dikes, cutting all the former rocks and consisting of an aplitic *soda-rich granite*. That they belong to the same eruptive series, with their source in the central part of the lens, is proved by the fact that they are confined to the norite field and occur in greatest quantity in the central field of quartz-norite. Here they form a network of narrow dikes, occupying about 7 per cent of the total area. The dikes are well defined, but with slightly blurred contacts. The same dikes occur also in the marginal norite, here only occupying about 3 per cent of the area, but more regularly and with razor-sharp contacts.

The very last products of volcanic action are some irregular veins of pegmatitic *potash-granite* and of snow-white, pure *quartz*, both carrying black tourmaline.

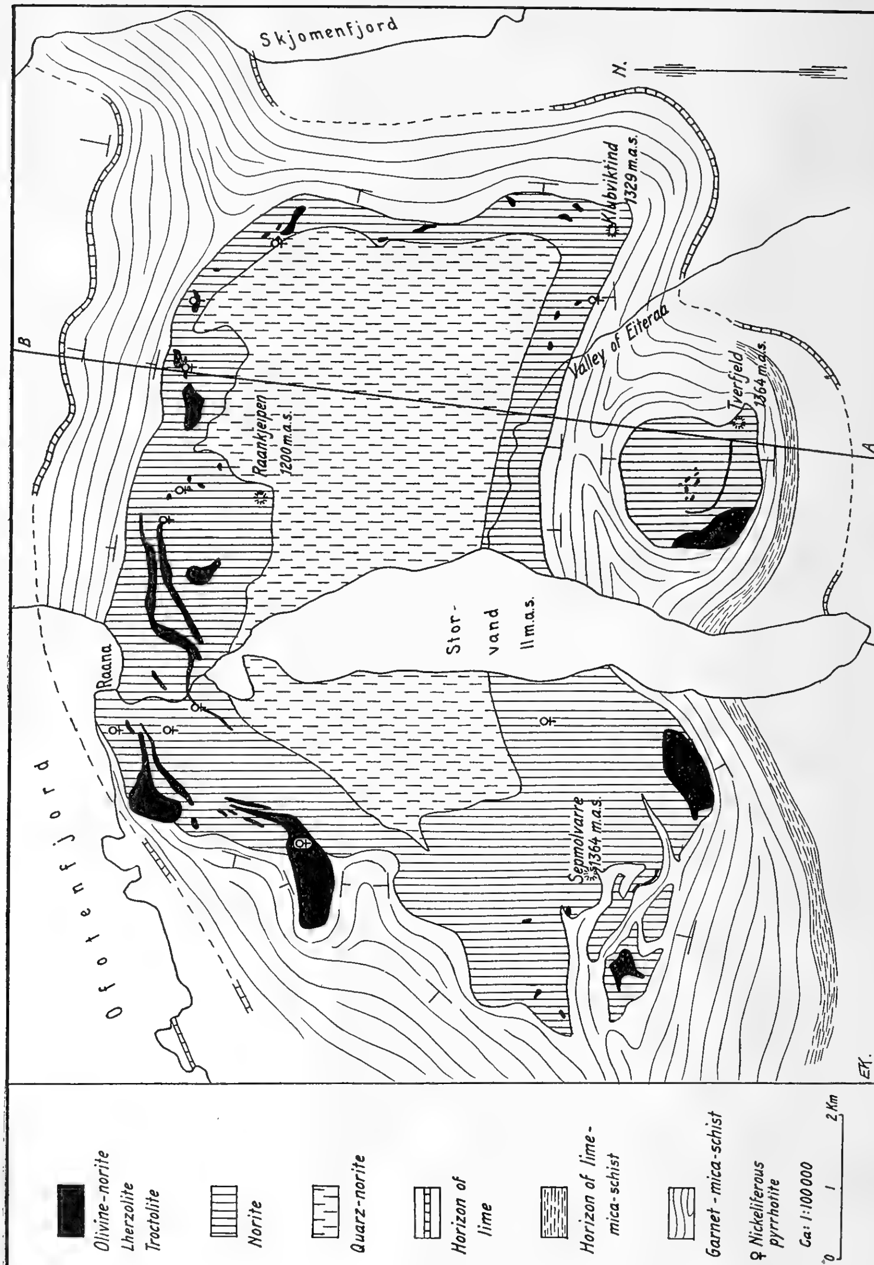

FIG. 1.—Map of the Raana norite field. Scale, 1:100,000

The distinct gap between the main mass of massive norites and the dike-formed acid differentiation products of aplitic granite reminds us very much of corresponding features of the "red rock" in the Duluth gabbro.[1]

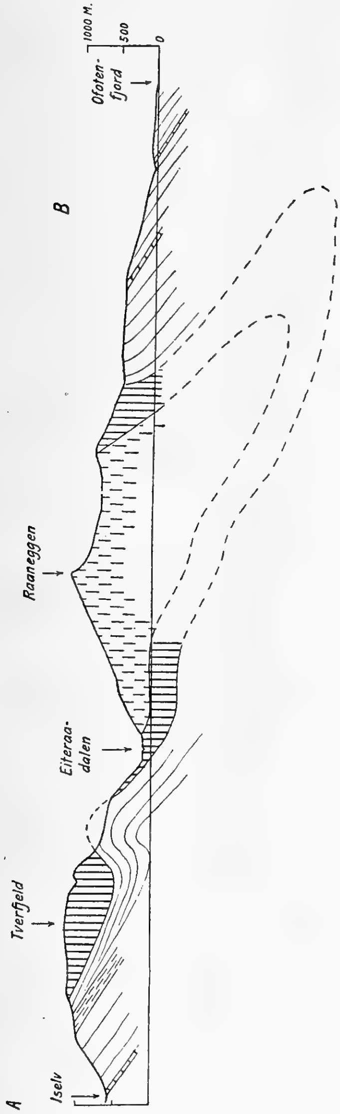

FIG. 2.—Section through the Raana field. Vertical and horizontal scale, 1:100,000

The deposits of nickeliferous pyrrhotite—which will not be treated here—are in their occurrence confined to the olivine-bearing rocks and to the marginal normal norite, and no traces of them are found in the quartz-norite or the younger dikes. In the olivine rocks the sulphides occur only as impregnations, but are relatively very rich in nickel. While the percentage of sulphur generally is between 1 and 2 per cent, the nickel content nevertheless reaches 0.7 per cent, of which not more than 0.1 per cent seems to be in the form of silicate in the olivine mineral. In the norite, the sulphides occur partly as impregnations, partly as segregated richer masses, but the nickel content in the pure sulphide is lower, generally between 1.5 and 4 per cent.

The mineralogical composition of the rocks is as follows:

The *olivine-bearing rocks* consist of olivine, rhombic and monoclinic pyroxene, and plagioclase, but generally no primary biotite. Accessory constituents are picotite, green spinel, magnetite, and pyrrhotite. The rhombic pyroxene is an enstatite with less than

[1] Frank F. Grout, "A Type of Igneous Differentiation," *Jour. Geol.*, Vol. XXVI, pp. 626 ff.

17 per cent ferrous silicate, the monoclinic pyroxene is a diallage. The plagioclase is always bytownite, about $Ab_{20}An_{80}$, nearly without zones, and occurring interstitially between the other minerals.

In the *normal norite* the minerals are: Hypersthene with abous 25 per cent ferrous silicate, poikilitic diallage in big individualt with hypersthene inclusions, labradorite-bytownite, often with zonal structure and varying composition, from 50 to 80 per cent An, and some primary biotite. Uralitization occurs in many parts of the field, but has no interest in this connection. The considerable variations in this rock seem exclusively to be due to quantitative variations in the relative proportion of pyroxene and plagioclase. The monomineralic rocks, anorthosite or pyroxenite, are however, never developed.

The *quartz-norite* contains iron-rich hypersthene with 40–45 per cent ferrous silicate, much diallage, which is here not poikilitic, but crystallized at the same time as the hypersthene, and further generally plenty of biotite. The plagioclase is labradorite with marked zonal structure, from 40 to 55 per cent An. Besides, this rock always contains small amounts of microcryptoperthitic orthoclase, rich in natron, and of free quartz.

The *aplitic granite* dikes contain quartz, oligoclase with 26 per cent An, non-perthitic microcline, and a little muscovite and biotite.

The *pegmatitic granite* dikes contain quartz, perthitic microcline with a faint green color, and black tourmaline. The perthite consists of 87 per cent microcline and 13 per cent albite-oligoclase with the composition $Ab_{83}An_{17}$.

The chemical composition of the differentiation products is seen from the analyses in the following table, where the calculated norm is also given.

The mode of these rocks naturally differs somewhat from the norm, even apart from secondary processes. So the potash, instead of forming orthoclase, partly enters into plagioclase and biotite, while part of the lime together with aluminia enters pyroxenes instead of plagioclase.

After an exact microscopic measurement, I give the mineral composition actually found in the normal norite and the quartz-norite, corresponding to the analyses No. 3 and No. 6 respectively.

ANALYSES

	(o)	1	2	3	4	(4a)	5	6	7
SiO_2	52.6	41.74	41.32	52.30	50.10	50.10	50.70	55.90	72.20
TiO_2	0.4	0.12	0.08	0.62		0.5	0.25	0.45	0.39
Al_2O_3	16.2	2.87	7.71	10.28		18.5	20.70	17.50	14.22
Fe_2O_3	0.5	2.96	1.83	0.37	26.75	0.5	0.50	0.08	0.55
FeO	6.7	10.66	8.63	8.80		6.3	4.38	6.97	2.00
MnO	0.1	0.35	0.13	0.20		0.10	0.07	0.09	0.03
MgO	12.2	38.70	35.26	18.30	12.54	12.54	9.43	5.98	1.03
CaO	7.7	1.48	2.22	6.06	8.65	8.65	10.35	8.32	2.60
BaO	Trace	....	None	Trace	....	....	Trace	Trace	0.02
Na_2O	1.8	0.63	0.54	1.14	....	1.5	1.58	2.75	2.55
K_2O	1.0	0.24	0.98	0.58	....	0.6	0.69	1.50	3.78
P_2O_5	0.05	0.09	0.06	0.05	....	0.05	Trace	0.05	0.08
Vd_2O_3	0.06?	....	....	....	0.09	0.09	....	....	....
S	0.06+	0.12	0.28	0.09	....	0.05	0.05	0.05	Trace
Ni	....	0.14	0.36	....	....	....	....	....	....
Cr_2O_3	0.06	0.05	0.10	....	0.09	0.09	0.08	None	....
Cl	....	....	....	....	....	....	....	....	Trace
F	....	....	....	....	....	....	....	....	Trace
CO_2	....	None	None	0.61	....	....	0.40	0.20	0.06
H_2O+	0.7+	0.25	0.39	0.86	....	0.7	1.12	0.50	0.65
H_2O-	....	(0.03)	(0.04)	0.06	....	....	0.07	0.07	0.06
Total	100.13	100.40	99.89	100.32	....	....	100.37	100.41	100.22
÷O for S	....	0.06	0.14	0.05	....	....	0.03	0.03	0
Total	....	100.34	99.75	100.27	....	....	100.34	100.38	100.22

CALCULATION OF THE NORM

	(o)	1	2	3	4	(4a)	5	6	7
Quartz	....	....	....	....	....	....	0.66	4.48	34.74
Orthoclase	6.12	1.39	6.12	3.34	....	3.34	3.89	8.88	22.24
Albite	15.20	5.24	3.99	9.43	....	12.58	13.62	23.23	21.48
Anorthite	33.08	4.31	11.12	21.41	....	41.98	47.26	30.88	12.79
Nefeline	....	....	0.39	....	....	....	....	....	....
Corundum	....	....	1.53	....	....	....	....	....	1.22
Σ sal	54.40	10.94	23.15	34.18	....	57.90	65.43	67.47	92.47
Diopside	2.44	1.65	....	3.56	....	0.89	1.33	7.20	....
Hypersthene	39.82	5.92	....	56.26	....	31.95	30.20	23.45	5.84
Olivine	0.62	76.52	72.72	1.98	....	6.88	....	....	....
Magnetite	0.70	4.41	2.55	0.46	....	0.70	0.70	0.12	0.70
Ilmenite	0.76	0.22	0.15	1.22	....	0.91	0.46	0.85	0.76
Apatite	....	0.34	....	0.12	....	....	....	0.12	....
Chromite	....	....	0.15	....	....	0.15	0.10	....	....
Pyrite	....	0.22	0.53	0.17	....	0.10	0.10	0.10	....
Σ fem	44.34	89.28	76.10	63.77	....	41.58	32.92	31.83	7.30
$MgO:FeO^{mol}$	3.5:1	7.3:1	7.3:1	4:1	....	....	4.3:1	1.6:1	1:1

KEY TO THE TABLE OF ANALYSES

No.	Petrographic Name	Symbol	Locality	Analyst
(o)	Calculated composition of undifferentiated magma	III.5.4.4	....	....
1	*Lherzolite*	(IV) V.1."5.1.(1) 2	Raanbog River	Naima Sahlbom
2	*Troctolite*	IV.1.5.1.1 (2)	Tverfjeldet	Naima Sahlbom
3	*Normal norite*	(III) IV.1.1.1.2	Source of Raanbog River	Olaf Röer
4	*Normal norite*	Partial analysis	Source of Raanbog River	Naima Sahlbom
(4a)	Calculated after 4	"III.5.4".4	....	....
5	*Uralitized norite*	II ".5.4" .4	Arnesskaret	Olaf Röer
6	*Quartz-norite*	II "." 5.(3) 4.4	Stemnes in Raana	Olaf Röer
7	*Aplitic granite*	I ".3 (4).2 (3).3	Sepmolvarre	Olaf Röer

While the composition of the quartz-norite is pretty uniform throughout, the marginal norite varies considerably. The analysis No. 3 represents a type very rich in hypersthene, analysis No. 5 a type rich in plagioclase, and analysis No. 4 a mean type.

	Normal Norite, No. 3 Weight Percentage	Quartz-Norite, No. 6 Weight Percentage
Quartz	None	2.1
Orthoclase	None	4.7
Plagioclase	20.0 (*ca.* $Ab_{40}An_{60}$)	59.2 ($Ab_{53}An_{47}$)
Diallage	4.6	16.6
Hypersthene	50.2	11.0
Biotite	7.3*	3.8*
Amphibole, secondary	16.1	Very little
Magnetite+ilmenite	1.4	2.4
Pyrrhotite	0.2	0.2
Rutile	0.2	None
Total	100.0	100.0

* A special case. Generally there is more biotite in the quartz-norite than in the normal norite.

By the aid of the seven analyses stated above we may calculate the mean composition of the four main groups of rocks occurring in the field. They are as follows:

	Olivine-bearing Rocks	Marginal Zone of Normal Norite	Central Field of Quartz-Norite	Aplitic Granite Dikes
SiO_2	41.53	50.80	55.90	72.20
TiO_2	0.10	0.44	0.45	0.39
Al_2O_3	5.29	16.99	17.50	14.22
Fe_2O_3	2.40	0.44	0.08	0.55
FeO	9.64	6.47	6.97	2.00
MnO	0.24	0.14	0.09	0.03
MgO	36.98	13.20	5.98	1.03
CaO	1.85	8.43	8.32	2.60
BaO	None	Trace	Trace	0.02
Na_2O	0.59	1.43	2.75	2.55
K_2O	0.49	0.60	1.50	3.78
P_2O_5	0.07	0.03	0.05	0.08
Vd_2O_3	....	0.09	....	....
S	0.20	0.06	0.05	Trace
Cr_2O_3	0.08	0.09	None	....

These variations are illustrated in the diagram of the differentiation processes (Fig. 3).

By measurement from the exact geological maps of the field we find the areal distribution of the main rock groups to be the following:

	km.²
Quartz norite	30.0
Normal norite	33.2
Olivine-bearing rocks	3.8
Total	67.0

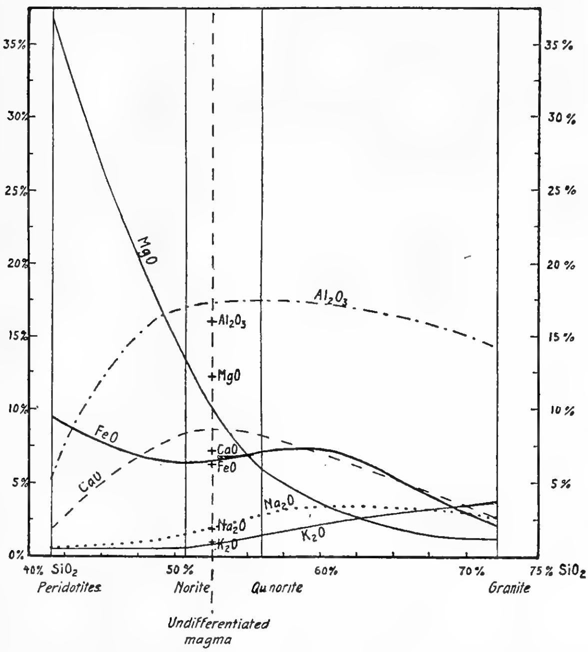

FIG. 3.—Differentiation diagram, based upon the mean composition of the main rock groups.

After the modifications necessary for the topographical irregularities, after recalculation to weight percentage, and after introducing the calculated quantity of the granite dikes, the proportion becomes the following:

	Weight Percentage
Aplitic granite	3.9
Quartz norite	31.3
Normal norite	57.4
Olivine-bearing rocks	7.4
Total	100.0

It may be noted here that these percentages are not proportional to the areal distribution of the rocks at the surface, because the quartz-norite and the normal norite without doubt are concentrically arranged throughout the field, so that any section through the center of the field will give the same image. Accordingly, their relative volume is with the greatest approximation calculated here after the formula for concentric spheres.

After the results given above, stating the quantitative proportion and the chemical composition of the main groups of rocks, we are in the rare and happy position to be able to state—with a fair amount of accuracy—the mean composition of the whole field, representing the original, undifferentiated eruptive magma. This composition is given in column 0, in the table of analyses. The most remarkable feature is the very high magnesia content in connection with the relatively high silica content. It is an ideal norite magma.

The regular, stepwise development of the differentiation products is best seen from the norm calculations. Not only the chemical composition changes along definite lines, but also the composition of the individual minerals, especially the plagioclase and the rhombic pyroxene. Very remarkable is the regular change in the ratio $MgO^{mol}:FeO^{mol}$ in the silicates, decreasing markedly toward the salic rocks. The reason for this appears from what follows.

THE DIFFERENTIATION PROCESS

Starting with a study of the known forsterite-diopside-silica and the forsterite-anorthite-silica diagrams,[1] we learn that in the mean, undifferentiated magma of the Raana field, nearly without olivine in the norm, olivine will anyway be the first mineral to crystallize—beginning at the border toward the surrounding schists—and it will continue to separate out until the eutectical line olivine-pyroxene is reached. By further cooling, rhombic pyroxene crystallizes out at the same time as the *already formed olivine crystals are resorbed.*

If this process continued undisturbed, the final result would be an olivine-free rock, corresponding to the norm. However, we find

[1] See N. L. Bowen, *Jour. Geol.*, Vol. XXIII, suppl., pp. 20 and 29.

in the field, as stated above, a series of olivine-rich rocks all around the marginal zone of the eruptive body and of nearly the same age as the environing norite. These olivine rocks can only have been individualized during that period of the crystallization process when free olivine crystals were really suspended in the crystallizing marginal zone. It is obvious that even the slightest accumulation of these crystals at certain places would at once bring them in excess, and prevent them from being wholly resorbed.

Consequently, olivine would become a part of the norm composition at those places. We may conclude, therefore, that such an accumulation of olivine crystals has really taken place. That gravity separation has not played a prominent part in it is obvious from the distribution of the olivine rocks. We are forced to conclude that convection currents and other movements in the magma have been able to bring about such an accumulation, possibly also a conglutination of already formed crystals.

The explanation is confirmed by two facts: First, in all the olivine rocks the olivine crystals are surrounded by resorption rims, proving that resorption has been going on, but stopped at a certain point. Secondly, the olivine crystals in all these fields have nearly the same proportion, $Mg_2SiO_4:Fe_2SiO_4$, quite independent of the quantity of olivine, and consequently have all separated out from a uniform magma.

We get a natural explanation of the varying quantities of plagioclase and pyroxenes in the olivine rocks and of their "swimming" character in the norite magma.

After the termination of the resorption period and the consolidation of the olivine rocks, the rest of the magma in the marginal zone should obviously be expected to have become poorer in magnesia and richer in silica than the magma of original composition in the still fluid central part. Field observations teach us, however, that the olivine-free norite in a very thick marginal zone has a more basic composition than the central part.

To explain this, we might think of compensating currents in the magma prior to the resorption period, which in combination with the resorption of some of the olivine might produce this basic

composition. Chemical calculations, however, show this to be impossible, and the differences in the composition of the plagioclase point in the same direction.

We might also consider another sort of differentiation, according to which the marginal zone of the massive represents the mean, original composition of the magma, intruded by younger femic and salic differentiation products. The evident field observations, however, contradict such a supposition also.

Here, therefore, the squeezing theory, as developed by N. L. Bowen (*loc. cit.*), turns out to be a very natural and obvious explanation.

It must be remembered that the process takes place under orogenic pressure. While the segregated crystals are only suspended in the magma, the pressure is static and has no influence on the differentiation. From the moment when the outmost shell forms a fixed crystal mesh, this shell has eventually to take up the mountain pressure, but, of course, at first is not able to do so. The pressure then will be dynamic. Following Bowen, this stage is supposed to occur when about 80 per cent of the mass is crystalline and only 20 per cent liquid.

As the volume of the magma is diminished by cooling and still more by crystallization, the outer shell will be compressed and its remaining interstitial liquid squeezed out. At that advanced stage of crystallization, this liquid will contain plagioclase, richer in soda, and pyroxene, richer in iron and lime than the already segregated crystals, moreover eventually potash-feldspar, free quartz, and magmatic water.

These components accordingly will move, and the direction of movement will be inward from the zone of dynamic into the zone of static pressure, because in the crystallizing zone with still static pressure the diminishment of volume will have the effect of releasing the pressure.

It is important to note that the process is here not supposed to be a squeezing for a long distance through a crystal mesh. Probably it is mainly a differential movement, restricted to the narrow transition zone, which moves inward at the same rate as the crystallization proceeds.

"The wave of crystallization is followed by a wave of squeezing." Continually the squeezed material will mix with the more fluid magma inside, which gradually becomes more acid.

The next step is the abrupt transition from the normal norite of the marginal zone to the quartz-norite of the central part. This is very easy to explain. It represents the stage when the outer, solid shell has grown sufficiently thick and strong to resist the compressing forces. From that moment no more compression and no more squeezing takes place. The remaining magma crystallizes quietly without further differentiation to a uniform quartz-norite, carrying a little free quartz and orthoclase, more acid plagioclase, and much biotite on account of enrichment of the magmatic water.

Of course, the contraction or the release of pressure continues as the crystallization proceeds, and finally results in a general formation of fissures in all directions, when the resulting stresses have grown sufficiently strong.

This fissuring obviously occurred at a stage when the quartz-norite was consolidated, with exception of the very last interstitial liquid, containing a considerable proportion of magmatic water and mineralizers. The liquid was drained into the fissures, forming aplitic dikes of soda-rich granite. This last separation is obviously not due to squeezing; but whether the liquid was really sucked out into the fissures or driven out by the pressure of the enriched gaseous mineralizers is difficult to tell. At any rate, there resulted a direct connection between these dikes and the last consolidated minerals in the quartz-norite, producing the slightly blurred contacts mentioned above.

Also in the marginal normal norite, clefts were formed at this period and filled with the same dikes of granite. Here they have razor-sharp contacts because the marginal rock at that time was completely solid.

The mineralizers once more separated out, carrying with them much potash and silica and giving rise to veins of pegmatite and pure quartz, both rich in tourmaline.

From the foregoing we see that the theory as modified covers all the observed facts. It is beyond doubt that the differen-

tation proceeded quite *in situ* and is confined to the crystallization period. Further we have learned that the squeezing differentiation gives results which in many respects are similar to those of gravitative separation. In the one case the liquid is removed from the crystals, in the other case the crystals from the liquid.

There is only one point left which cannot yet be explained in detail: how some of the primary olivine crystals could be brought to accumulate or conglutinate at certain places. On this point, therefore, we can only state the fact and remark that any supposition of immiscibility of olivine with the rest of the magma, resulting in the formation of fluid drops of olivine, would not help us, but be in opposition to several of the above-stated facts.

As will easily be seen, the special differentiation type of Raana is not apt to occur very frequently. It presupposes the following conditions: (1) the crystallization must take place under lateral pressure; (2) it must take place completely *in situ*, in a closed room without supply of new magma during the process; (3) gravitative separation must not play a prominent part.

In the same district we meet a somewhat different type of squeezing differentiation, one which will prove to have a more general occurrence in orogenic folding zones. It occurs when the central part of the eruptive mass is not—as in Raana—protected against lateral pressure. A short statement of it is here added.

THE AMPHIBOLITE SERIES

As mentioned in the first part of this paper, in the folded mountain region there occur considerable quantities of femic eruptive rocks very intimately injected into the schists and completely recrystallized through dynamic metamorphism. They do not occur in so extensive individual masses as do the fresher rocks of the Raana type, but they are much more widely distributed. Chemically they have diabasic composition and differ from the corresponding rocks of the Raana field in carrying more iron in proportion to magnesia, more sodium, titanium, and phosphorus.

On account of the total parallelism of all eruptive injections in the district, direct observations of relative ages can generally not

be made. The amphibolites *might* be somewhat older than the Raana norites and accordingly have been subject to a longer period of dynamic metamorphism during the Caledonian folding. But the reason for the different degree of metamorphism might also be that the process has been able to act more severely upon these rocks on account of their lesser thickness and their very intimate injection in the surrounding schists—which show about the same degree of metamorphism—while the massives of the Raana type have resisted better.

The mineral composition of the ordinary amphibolites is: amphibole and acid plagioclase as the predominant minerals, more or less epidote or clinozoisite, quartz, leucoxene, and often garnet and biotite. None of the primary minerals are left.

The original basic plagioclase has been more or less albitized, producing plagioclases from albite to oligoclase composition, while part of the lime enters the epidote minerals. The original pyroxenes have been changed to amphibole, and it is noteworthy that while in the Raana field the uralitization has produced a nearly colorless amphibole of actinolitic composition, poor in aluminia, the metamorphic rocks contain common green amphibole, rich in aluminia.

The differentiation shows several features analogous to those just described from the Raana field, but also significant differences.

In many of the amphibolite zones we find a number of small bosses of serpentine rocks, very irregularly distributed. The dozens of such bosses observed nearly always occur within the amphibolite rock and are not separately injected into the schists. They are all of relatively small dimensions, rounded or lens-shaped, and sharply defined from the surrounding amphibolite. They obviously correspond to the peridotitic rocks of the Raana field, but are always completely metamorphosed to serpentine and talc minerals.

In numerous cases the amphibolite itself is nearly homogeneous, and without intermediate steps there is a wide gap over to soda-rich granites which occur nearly everywhere in intimate connection with the amphibolite series in such a way that there can be no doubt of their mutual relation as differentiation products.

Here the granite does not occur as crossing dikes, but as small lenses or bands arranged parallel to the schistosity of the amphibolites. In some cases they occur as separate sheets of considerable thickness in or at the border of the amphibolite. Only in rare cases, when the amphibolite has retained a more massive structure without marked schistosity, do they occur as numerous crossing, irregular stringers cutting the femic rock.

The contrast between the two rocks is still more marked than was the case in the Raana field on account of the great difference in chemical composition and the different degree of metamorphism. While the amphibolites are completely recrystallized, the granitic differentiates have retained much of their original structure, partly because they are somewhat younger, partly because the granitic mineral association has by far not the same tendency toward mineral readjustment under new conditions as is the case with the femic rocks.

These granites are characterized as soda-rich, but their composition may differ somewhat. Sodium may be quite predominant among the alkalies, the rock becoming a *Trondhjemite* as described by V. M. Goldschmidt from the Trondhjem district farther south. By increasing potash they get a more granodioritic composition, up to a limit with about equal molecular amounts of soda and potash. In all cases they are poor in dark minerals and generally have aplitic structure.

They are distinctly different from the ordinary granites in the district, which occur as great independent eruptions, and where the potash is always predominant.

The chemical composition of the ordinary amphibolite and its granitic differentiate is seen from the analyses on page 719.

In some cases we meet intermediate rocks between the amphibolitic and granitic extremities, mainly of dioritic composition. The variations are never regular, but form "schlieric" or banded alternations in the schistose rock.

From the foregoing we have seen that the characteristic regular differentiation step between the normal norite and the quartz-norite in the Raana field has no correspondent in the amphibolite series. This is a natural consequence of the fact that here the

central part of the mass was not protected against the lateral pressure which prevailed during the whole crystallization.

The result of the differential squeezing during the crystallization period has therefore been the ultimate separation of the last consolidating constituents of the magma. During the continued pressure after crystallization of the main rock mass, this last residue has been squeezed into the already consolidated rock as

ANALYSES

	No. 8	No. 9
SiO_2	49.12	67.49
TiO_2	2.46	0.29
Al_2O_3	13.70	16.52
Fe_2O_3	1.95	2.33
FeO	10.67	0.34
MnO	0.18	0.03
MgO	5.19	0.88
CaO	11.02	3.43
BaO	Trace	Trace
Na_2O	2.58	5.76
K_2O	0.44	1.93
P_2O_5	0.22	0.50
S	0.19	0.10
CO_2	1.21	None
H_2O+	1.37	0.81
$H_2O\div$	0.01	(0.05)
Total	100.31	100.41
÷O for S	0.10	0.05
Total	100.21	100.36

CALCULATION OF THE NORM

	No. 8	No. 9
Quartz	3.36	19.74
Orthoclase	2.22	11.12
Albite	22.01	48.73
Anorthite	24.46	13.34
Corundum	....	0.10
Σ sal	52.05	93.03
Diopside	17.51	....
Hypersthene	18.17	2.20
Hematite	....	2.24
Magnetite	2.78	....
Ilmenite	4.71	0.61
Apatite	0.52	1.34
Pyrite	0.36	0.19
Σ fem	44.05	6.58
$MgO^{mol}:FeO^{mol}$	1.24:1	....

KEY TO THE TABLE OF ANALYSES

No.	Petrographic Name	Symbol	Locality	Analyst
8	*Garnet-amphibolite*	III.″5.″4.″5	Björkaasen Mine	Olaf Röer
9	*Aplitic granite*	I″.4.2.4″	Brugsaas	Naima Sahlbom

lenses, stringers, and bands of aplitic granite. Their ordinary arrangement parallel to the schistosity shows that this must have been partly developed already during the consolidation as crystallization schistosity.

The magmatic water and mineralizers with their dissolved substances seem to have taken their own path, having a more active power of motion contrary to the purely passive motion of the ordinary squeezed material.

They have given rise to replacement phenomena and formation of ore deposits, still younger than the aplitic granite.

GEOLOGIC RECONNAISSANCE IN BAJA CALIFORNIA

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During 1921, I spent several months in Baja California determining geologic relations of a large area, and as there is but little on record regarding the geology of this region, it is desirable to set forth such of my observations as appear to be of general interest. I journeyed with pack train from near Ensenada southward to beyond Comondu and later extended the reconnaissance to beyond La Paz as far as Todos Santos. Most of the observations were in the district between the high sierra and the west coast, but the peninsula was cross-sectioned near San Ignacio and Santa Rosalia, from Mulegé to La Purisima, and in the La Paz region (Fig. 1).

Most of my attention was given to the sedimentary rocks of Cretaceous and Tertiary age, but the limits and character of some of the crystalline rocks were ascertained. Many of the general relations are set forth in the cross-sections of Figures 2, 3, and 4.

PRE-CRETACEOUS

The lowest strata observed were schists and other metamorphic rocks, invaded by white, massive, coarse-grained granite. At most places this granite includes fragments of schist, and many clearly intrusive contacts were noted. The granite constitutes the high Sierra San Pedro Martir in the north central part of the peninsula, where one or two summits reach an altitude of about 10,000 feet. It also outcrops north of Ensenada, north of Santa Caterina, and in many high central ridges from latitude 29° 30′ to latitude 28°, and is prominent in the sierra at the lower end of the peninsula southeast of La Paz. I crossed the schists in several broad zones between Santa Caterina and Calmalli and noticed that a few miles

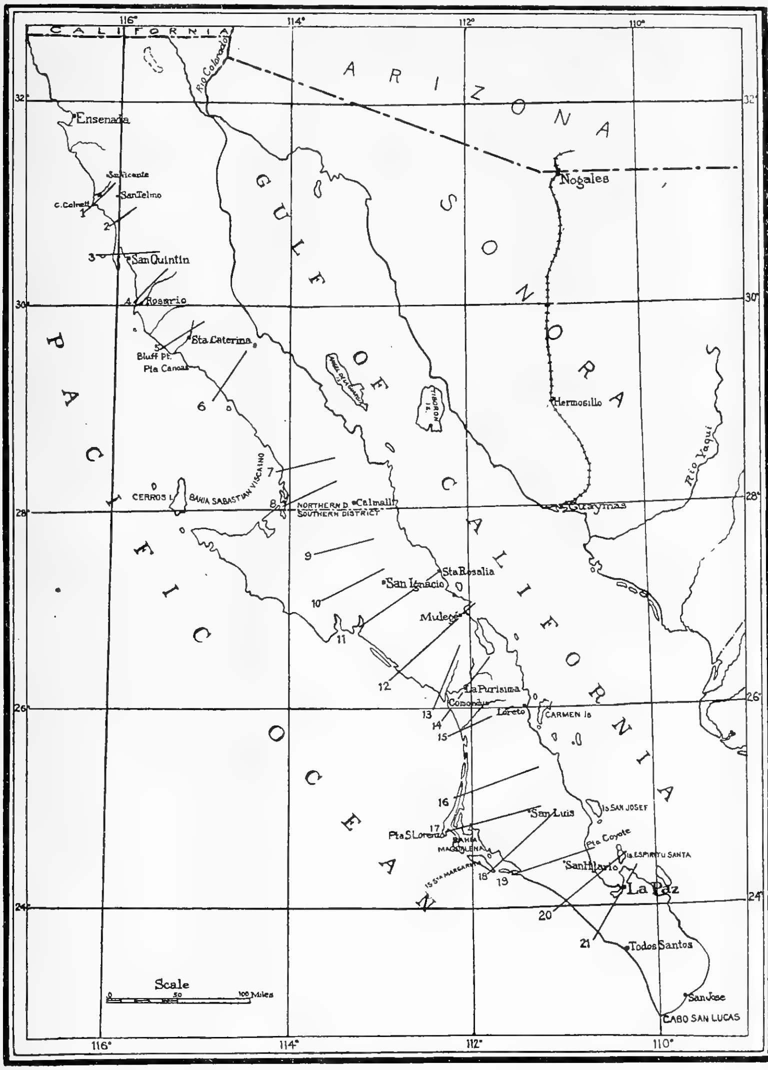
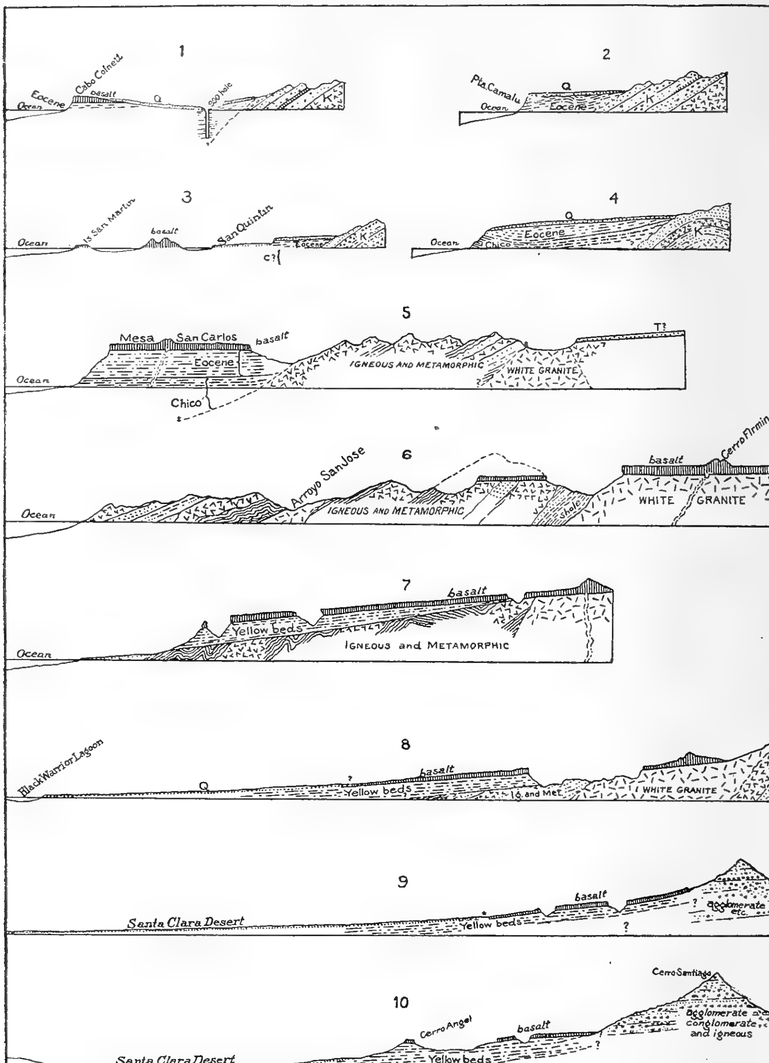


FIG. 1.—Map of Baja California showing principal shore features and settlements, and location of cross-sections in Figures 2, 3, and 4.

south of the latter place they are overlapped by younger sediments. They appear again in high ridges in Cerros Island and from Punta San Lorenzo to Isla Santa Margarita on the west side of Bahia



Scale: Hor. 5 miles Vert. 3000 feet

FIG. 2.—Sections across parts of northern and central Baja California
Q=Quaternary

de la Magdalena.[1] They also occur in the south end of the peninsula in the Triunfo mining district south of La Paz.

[1] According to J. Ross Brown, *Resources of the Pacific Slope*, San Francisco, 1869, p. 143.

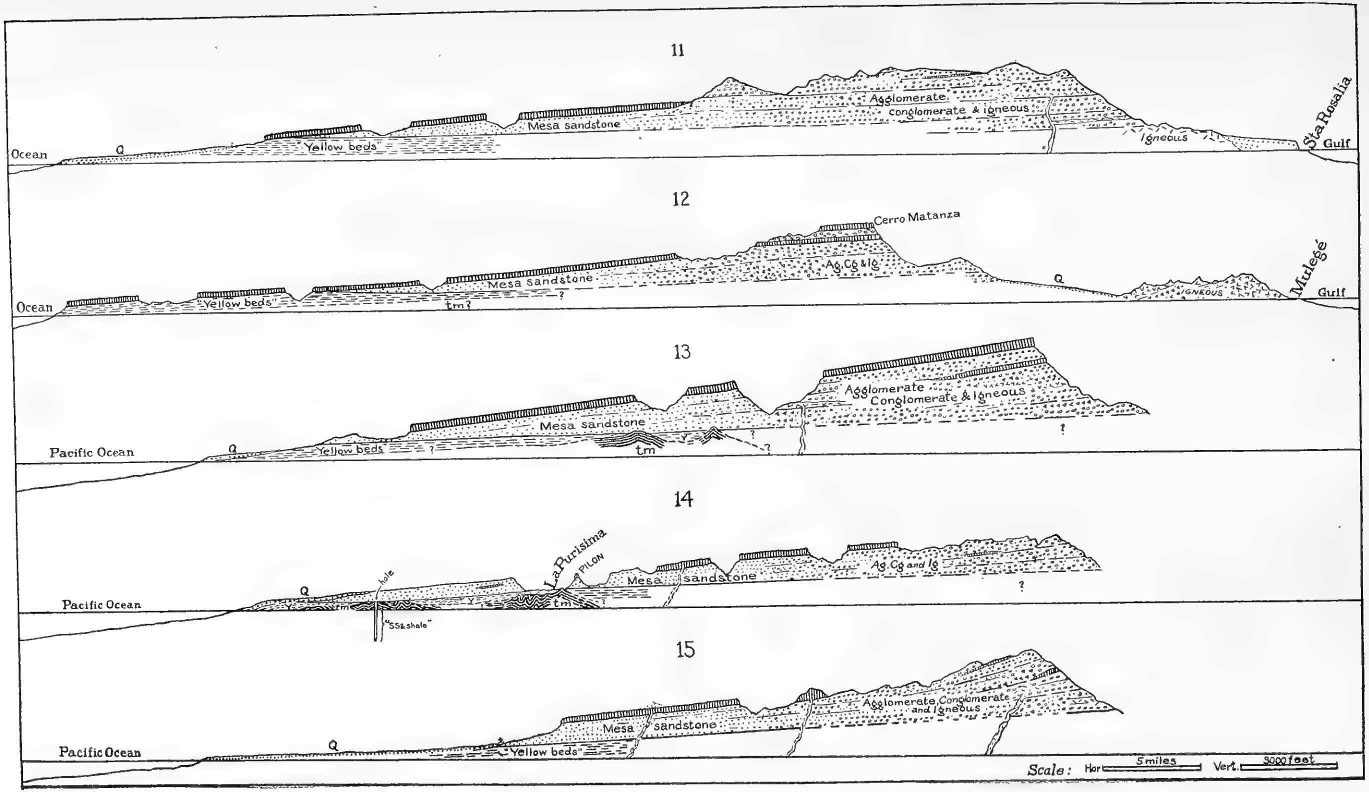


FIG. 3.—Sections across central Baja California
Q=Quaternary

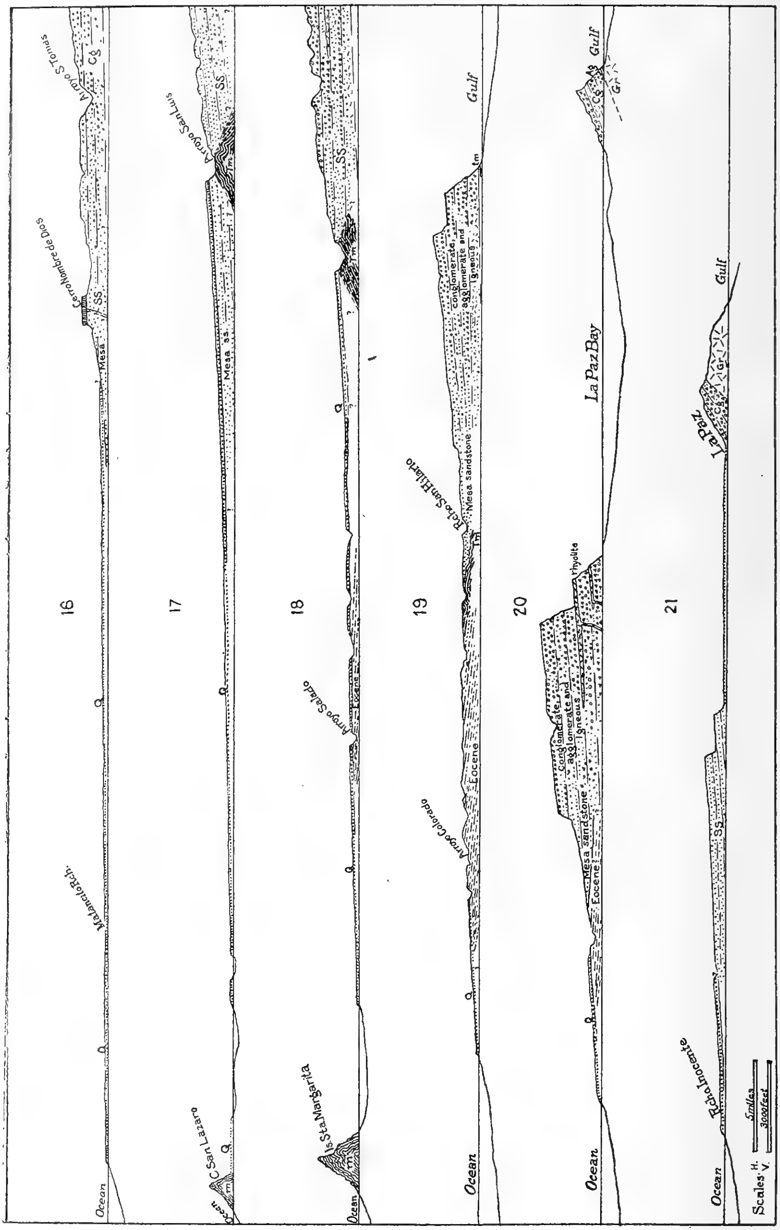


FIG. 4.—Sections across part of southern Baja California

Lindgren[1] has given considerable information regarding schist and granite in the northern part of the peninsula; Gabb[2] noted some features observed on his trip north from La Paz; and Emmons and Merrill[3] described relations of the schists near the onyx quarries east of Santa Caterina.

I obtained no evidence as to the age of these granites and schists, but my general impression was that they were much older than Cretaceous. I did not observe any overlap by Cretaceous rocks, but I saw Tertiary deposits lying on or against them at various places. In the region east of La Paz, the granite appears to be the floor on which the Tertiary volcanic series lies. This relation is strongly indicated at Cabo Lobos on the east side of Isla Espiritu Santo where a separating fault appears hardly possible. A similar relation is indicated in Sierra Giganta, northwest of Loreto. The structural relations of the ridges of schists in Isla Santa Margarita and extending northward to Cabo San Lazaro is not known, and it is not unlikely that these rocks are the uplifted floor of the strata farther east.

CRETACEOUS SYSTEM

General relations.—A large part of the peninsula is underlain by Cretaceous rocks which outcrop extensively in the central and northern parts. Two principal series are present, both of late Cretaceous age and separated by an uncomformity, the older series, of unknown correlation, having been uplifted, flexed, and cut by large igneous masses before the younger series, Chico, was deposited. The Chico beds were not observed in contact with the older series, and the principal evidence of the separateness of the two is the difference in attitude of the beds throughout the area and the fact that the older series is uncomformably overlapped by

[1] Lindgren, "Notes on the Geology of Baja California, Mexico," *Proc. Cal. Acad. Sci.*, Second Series, Vol. I (1889), pp. 173–96, and Vol. II (1890), pp. 1–17; and "Geology and Petrography of Baja California, Mexico," *Proc. Cal. Acad. Sci.*, Second Series, Vol. III (1890), p. 26.

[2] W. B. Gabb, *Geol. Survey of Cal.: Reports*, Vol II (1869), Appendix, pp. 1–20.

[3] S. F. Emmons and G. P. Merrill, "Geological Sketch of Lower California," *Bull. Geol. Soc. Am.*, Vol. V (1894), pp. 489–514.

the Eocene strata which lie with apparent conformity on the Chico beds. This relation is shown in the following section (Fig. 5).

These pre-Chico Cretaceous rocks consist of conglomerates, quartzites, tuffs, and agglomerates with large bodies of interbedded eruptive rocks. They are also cut by dikes and large stocks of igneous rocks of various kinds. In many localities theigneous rocks predominate over the sediments or pyroclastics, and in places there is much metamorphism. Unaltered or but little altered sandstone and shales appear in places, notably near old San Domingo Mission, 25 miles north of San Quintin, where they contain many large oyster shells, and in the Arroyo San José, 40 miles southeast of Santa Caterina. Limestone also occurs. It is conspicuous north and northeast of the ruins of Mission San Fernando, 30 miles due east of Rosario, where the relations shown in the following sketch section (Fig. 6) are presented. The limestone is filled with fossil oysters of upper Cretaceous age. Part of the outcrop is shown in Figure 7.

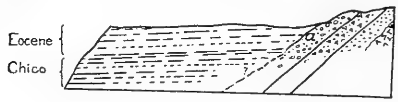


FIG. 5.—Sketch section showing relations of Eocene and Chico beds to metamorphic and igneous rocks of Cretaceous (pre-Chico) age, near latitude 30°, west coast, Baja California. *a*, exposed overlap of Eocene beds.

These metamorphic Cretaceous rocks appear to extend as far south as latitude 28° 40′, and possibly the schists and other rocks,

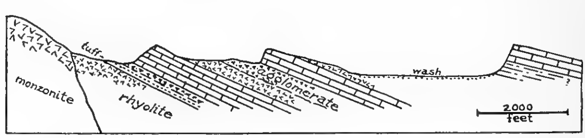


FIG. 6.—Sketch section showing relations of Cretaceous limestone north of the ruins of San Fernando Mission, 30 miles east of Rosario, Baja California.

although apparently of an older series and extending to Calmalli, may be metamorphosed Cretaceous sediments. Apparently they are so regarded by Bosé and Wittish in their brief statements regarding relations in the Northern District.[1]

[1] "Memoria de la Comisión del Instituto Geológico de México que exploró la región Norte de la Baja California," *Inst. Geol. de México, Perergones*, Vol. XIV (1913), pp. 307–533.

CHICO FORMATION

The upper part of the Chico formation (late Cretaceous) rises above sea-level a few miles north of Rosario, and it remains in view along the ocean bluffs and lower parts of the valleys of the Rio Rosario, Arroyo San Vicente, Rio San Fernando, and Arroyo Santa Caterina as far as latitude 29° 24′, a few miles southeast of Punta Canoas. The rocks are soft sandstone and shale of light-gray to buff color with round concretions at most places. Emmons and Merrill[1] noted the occurrence of Upper Cretaceous fossils at a point about 3 miles north of Santa Caterina Landing. Specimens collected from massive sandstone near sea-level and in calcareous layers 200 feet above were determined by T. W. Stanton as follows: *Arca breweriana* Gabb; *Baculites chicoensis* Trask; *Tessarolax distorta* Gabb; and *Inocerami*, not determinable.

FIG. 7.—Fossiliferous Cretaceous sandstone northeast of San Fernando Mission, twenty miles east of San Rosario, Baja California.

The fossils which I obtained in these beds at the Arroyo Hondo, 15 miles north of Rosario, were identified by Dr. T. W. Stanton as follows: *Ryhnchonella* sp.; *Ostrea* sp.; *Inoceramus whitneyi* Gabb; *Yoldia nasuta;* *Nemodon vancouverensis* Meek?; *Dentalium* sp.; *Gyrodes* sp.; *Anchura* sp.: *Cinulia obliqua* Gabb; *Baculites chicoensis* Trask; and *Baculites occidentalis* Meek. All these fossils are typical of the Pacific Coast Chico which is of Upper Cretaceous age.

Lindgren[2] has reported the occurrence of the Chico fossil, *Coralliochama orcutti*, in sandstones appearing in a small area among the volcanic rocks a few miles southwest of Ensenada.

[1] *Op. cit.*, p. 501.

[2] *Op. cit.*, p. 176.

TERTIARY SYSTEM

EOCENE

General relations.—I found that on the shore of the Pacific Ocean the earlier Tertiary deposits began at the mouth of the Rio San Vicente in latitude 31° 30′ and extended south to about latitude 29° 25′, south of Punta Canoas. In this area they underlie a narrow coastal plain mostly from 8 to 10 miles wide, but broadening to nearly 20 miles near latitude 30°. They lie on the Chico beds but to the east abut against the metamorphic pre-Chico rocks, and the termination of the crop to the north and south is due to the westward extension of the latter to the ocean shore. Throughout the area the strata dip at very low angles to the west, and local flexures are rare and slight. Some general relations are shown in sections 1–5, Figure 2.

The rocks are shales and sandstones mostly of light-gray to buff color, 1,200 feet or more thick near latitude 30°. To the south they contain fossils of the Martinez or Middle to Lower Eocene age, but in the region north of latitude 30° most if not all of the beds appear to be Tejon or Upper Eocene, although the lower formations may exist beneath the surface. There are prominent exposures along the lower parts of the valleys of the Rio Rosario, Arroyo San Vicente, Rio San Fernando, and Arroyo Santa Caterina, near Rosario and southward, and in bluffs along the ocean near Cabo Colnett and Punta Camalu and near Bluff Point and Punta Canoas, the two latter southwest of Santa Caterina. As shown in section 5, Figure 2, the thick succession of beds exposed in the bluffs along the ocean at Bluff Point is preserved by the heavy lava cap of Mesa San Carlos.

Fossils.—Fossils were collected at several localitities in the Eocene beds and kindly determined for me by Dr. Julia A. Gardner. On the shore of the Pacific Ocean, a mile south of the mouth of the Arroyo San Antonio, latitude 31° 05′, the following were found: *Cylichna* sp.; *Turritella* n.sp., cf. *T. uvasana* Conrad; *Amauropsis* sp.; *Leda* sp.; *Cucullaea matthewsoni*, Gabb ?; *Cardium*, cf. *C. breweri* Gabb; *Cardium* sp.; *Tellina?* sp.; and *Semele* sp.

On the ocean shore a half-mile south of Colnett Creek, 5 miles southeast of Cabo Colnett, were collected *Omphalius* sp.; *Cardium*

n.sp., cf. *C. quadrigenarium* Conrad. In the bluffs 5 miles east of San Quintin there were collected *Turritella* n.sp.; *Leda* sp.; *Tellina* sp., cf. *T. hornii* Gabb; and *Spisula* sp., cf. *S. merriami* Gabb. In the drift at the mouth of San Simon Cañon, 6 miles southeast of San Quintin, *Cardium cooperi* Gabb ? was collected. These are all regarded as Tejon or Upper Eocene in age. The following fossils were collected from two horizons in the slopes about 5 miles northeast of Santa Caterina Landing: Upper beds, *Surcula* sp.; *Heterotoma gabbi* Stanton: *Turritella pacheococensis* Stanton ?; *Turritella* sp.; *Natica (gyrodes)*, cf. *N. lineata* Dickerson; *Amauropsis* sp.; *Dentalium cooperi* Gabb ?; *Leda* sp. 2; *Glycimeris* n.sp. aff. *G. veatchii* var. *major* Stanton; *Cucullaea matthewsoni* Gabb; *Ostrea* sp. ind.; *Pecten* sp.; *Pinna* sp.; *Crassitellites* sp. ?; *Venericardia planicosta* Lamarck var.; *Phacoides* sp., cf. *P. diegoensis* (Dickerson); *Phacoides* sp.; *Cardium cooperi* Gabb ?; *Cytherea* sp. aff. *C. hornii* Gabb; *Spisula* sp. The age of these is regarded as Upper Martinez or Middle Eocene.

From beds 200 feet lower were collected: *Cucullaea matthewsoni* Gabb ?; *Ostrea* sp. ind.; *Anomia* sp.; *Lima multiradiata* Gabb; *Phacoides* sp. ind. The age of these is regarded as Lower or Middle Martinez or Lower Eocene.

Emmons and Merrill[1] reported fossils from rolled pebbles of impure limestone obtained along the beach of the south of Santa Caterina Landing "which had evidently fallen from the cliffs above, and from a bed of similar composition in place at what was assumed to be about 1,200 feet higher in horizon, at San Carlos anchorage (collected by Mr. A. E. Foote), 8 miles north of Bluff Point." These were determined by Dr. T. W. Stanton as follows: *Cardita planicostata* Lam., *Leda gabbi* Conrad, *Urosyca caudita* Gabb, and undetermined species of *Nucula*, *Pectunculus*, *Tellina*, *Turritella*, *Dentalium*, and *Crassatella*. These forms were regarded as Tejon (Upper Eocene).

EOCENE (?) WEST OF LA PAZ

Sandstones which appear to be of Eocene age outcrop in an oval area of about 200 square miles in the valleys of arroyos Liebres, Colorado, San Hilario, and Guadalupe, all west of San Hilario.

[1] *Op. cit.*, pp. 501–2.

Smaller exposures are revealed by the Arroyo Salado and its branches near Rancho Sauce, 35 miles west-northwest of San Hilario, and by arroyos Conejo and Datilari, 20 miles south-southeast of San Hilario (or about 40 miles due west of La Paz).

The rocks are light-gray sandstone, mostly soft, but with harder layers and hard concretions. Some argillaceous members are included and some of the sandstone layers have a greenish tint. In the extensive exposures on the Arroyo Colorado about Rancho Tepetate where dips are from 3° to 5°, the thickness is not less than 3,000 feet unless the strata are duplicated by faulting. The dips are somewhat steeper near Rancho Sauce where more than 2,000 feet of beds are exposed. One of the largest outcrops in this vicinity is just south of Rancho Santa Rosa where sandstone ledges extend along the banks of the arroyo for several miles. It is capped by late Tertiary or Quarternary conglomerate and limestone. In the valley of the Arroyo de los Liebres, ledges of sandstone and sandy clays are exposed in a wide area of low ledges and buttes which extend to, or nearly to, the mouth of the arroyo. All the dips are at low angles to the north.

FIG. 8.—Sandstone (Eocene?) on the Arroyo Colorado one mile below Rancho Tepetate, ten miles west of San Hilario, Baja California.

The exposures in the valley of the Arroyo Colorado extend from the junction with the Arroyo de los Liebres nearly to the mouth of the Arroyo Caracol. Prominent cliffs of the sandstone occur at

Rancho Tepetate and for some distance above and below that place (Fig. 8). Dips are all to the north or north-northeast at low angles. The outcrop of the formation extends up the Arroyo San Hilario to within about 4 miles of Rancho San Hilario where the top member is massive, pale buff sandstone with irony layers and many fragments of oyster shells. It dips north and apparently passes beneath the Monterey beds. There sandstones have yielded few fossils, but echinoid spines and foraminifera are abundant at Rancho Tepetate, and oysters and a shark's tooth near Rancho Santa Teresa all appear to indicate Eocene age. Evidently the exposures are due to a mound of the formation extending southeast and southwest with altitude sufficiently great in the region from the Arroyo Salado to the Arroyo Conejo for the strata to be revealed by erosion in the deeper valleys.

MIOCENE

Monterey beds.—In the vicinity of the oases of La Purisima and San Hilario I found small exposures of strata so closely resembling the Monterey formation of southern California that tentative correlation seems desirable. Outcrops extend along the Arroyo de la Purisima from 2 miles above tidewater to within 6 miles of La Purisima and a small one appears in the upper part of the latter village. Outcrops also extend along the valley of the Arroyo San Gregorio for a mile or more about 10 miles southwest of La Purisima, and smaller crops appear above Purisima Vieja and San José, respectively 15 and 20 miles northwest of La Purisima, and on the Arroyo San Raimondi, 35 miles northwest of La Purisima. The same beds appear in an area of 2 or 3 square miles a short distance west of San Hilario, and in small crops of steeply upturned beds appearing at intervals from a point a mile northwest of San Luis to a point 15 miles southeast of that place. The relations are shown in sections 13, 14, 17, 18, and 19, Figures 3 and 4. At most places the beds are more or less tilted and flexed, and various younger formations overlie them unconformably, as shown in Figures 9 and 10, although at several places where the strata are not flexed there appears to be gradation into the overlying "yellow beds."

The principal rocks of the Monterey beds are gray to pale-buff, fine-grained sandstone and sandy shale with abundant fish scales. They include interbedded diatomaceous deposits mostly from 1 to 3 feet thick and some thin layers of glassy silica from buff to black in color. More than 500 feet of beds are exposed along the Arroyo de la Purisima where for several miles the formation extends from 5 to 20 feet above the bed of the arroyo and there are many flexures.

The southernmost exposures in the Arroyo de la Purisima in the west end of the Big Bend about 2 miles above tidewater contain much fine, white diatomaceous earth interbedded in fine-grained, yellowish sandstone and compact shale with many fish scales. In

FIG. 9.—Flexures in the Monterey beds on the Arroyo de la Purisima, ten miles below La Purisima, Baja California.

this region the dips are to the west or southwest, so that higher beds rise as the valley is ascended. In about 3 miles the lowermost beds appear in the crest of a low anticline, and from these were obtained the following fossils, determined by Dr. Julia A. Gardner: *Scutella andersoni*; *Chrysodomus* sp.; *Turritella* sp., cf. *T. margaritana* Normand; *Vermetus* sp.; *Leda* sp.; *Arca*, cf. *A. medio-impressa* Clark; *Pecten*, cf. *P. crassicardo* Conrad; *Crassitellites* (*Crassinella*) sp.; *Cardium* sp. nov.; *Chione* sp.; *Cytherea* sp.; *Mecoma* sp.; *Solen* sp.; and some corals and bryozoa. The age is regarded as probably Vaqueros, equivalent to Lower Monterey. These lowest beds appear again in another small uplift a short distance north, beyond which the strata descend in apparent regular order showing extensive exposures in the banks of the arroyo as shown in Figure 9. Here the formation is overlain by conglomerate.

Yellow beds.—Under this title I shall group deposits of late Miocene age exposed extensively along the western slope of the peninsula from latitude 29° to latitude 24°. They are mostly soft, loamy sandstone and sandy clay of pale straw-yellow tint with local limy beds, the latter generally full of fossils. Some contemporaneous igneous rocks are included in places. The yellow beds have an aggregate thickness of 500 feet near the coast between San Ignacio and La Purisima, although at the latter place the amount is not more than 120 feet. As stated above, there is unconformity between this formation and the Monterey beds, only noticeable, however, where the latter are flexed; the overlying "mesa sandstone" is also unconformable, as shown in Figures 11 and 12.

FIG. 10.—Steeply tilted Monterey beds on east side of the Arroyo de la Purisima, six miles below La Purisima, Baja California.

The yellow beds are first noticeable in thin, scattered bodies lying on the metamorphic series near latitude 29°. Near latitude 28° 30′ they have the relations shown in section 7, Figure 2, having been preserved from erosion by a thick cap of basalt. This relation continues for many miles south, as shown in sections 8–15, although the floor of metamorphic rocks sinks out of sight a short distance south of Calmalli. In the La Purisima region and near San Hilario the underlying formation is the rather uneven surface of the Monterey beds. Probably the formation thickens considerably under the Santa Clara Desert, where doubtless it is underlain

by later Cretaceous rocks such as those reported to appear in the southern portion of the Santa Clara Mountains.

Local features.—About 150 feet of the yellow beds are exposed under the lava cap at Cerro Angel, 17 miles west of San Ignacio (see sec. 10, Fig. 2). The principal material is gray to pale greenish-yellow sandstone with beds of volcanic ash. There are extensive exposures about San Ignacio where the strata are capped by basalt, as shown in Figures 13 and 14.

FIG. 11.—Upturned "yellow beds" overlain unconformably by Mesa sandstone, six miles west of La Purisima, Baja California.

FIG. 12.—"Yellow beds" overlain unconformably by Mesa sandstone on Arroyo de la Purisima, five miles below La Purisima, Baja California.

The yellow beds in the exposures extending from the Arroyo Valle to beyond Rancho Quarente, about 40 miles south of San Ignacio, present a uniform succession, about 500 feet thick, of soft, yellowish sandstone and loam. The base is not exposed and the top is eroded. Some beds contain considerable clay, and others are nearly pure sand. The principal color is a pale greenish-yellow. A few thin layers of hard sandstone and conglomerate are included. A hard fossiliferous layer occurs in places in the middle of the beds exposed.

The yellow beds outcrop for many miles along the Arroyo San Raimondi (or San Miguel), 30 miles northwest of La Purisima, from a point a half-mile above Rancho las Tules to its mouth, with relations shown in Figure 16. Pale-yellow, loamy sands or soft sandstone prevail, and at several points, notably at Rancho San Antonio and near the outcrop of the Monetrey beds 6 miles southwest of Caije (a very small settlement 35 miles northwest of La Purisima), a basal limestone member is exposed filled with

FIG. 13.—San Ignacio, Baja California, from the south. High sierra in distance; later Miocene capped by basalt in middle distance.

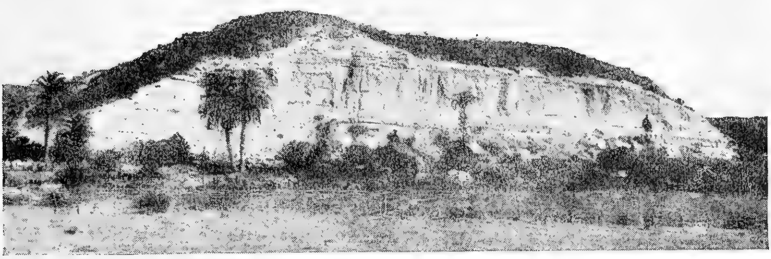


FIG. 14.—Upper Miocene strata capped by basalt at San Ignacio, Baja California

fossils (not determined). Agglomerate and sheets of lava occur in the lower part of the formation in this valley, as shown in Figure 16.

In the valleys of the Arroyo Juanico and the Arroyo Mesquital, 20 to 25 miles west of La Purisima, the yellow beds are exposed in slopes and cliffs of considerable extent, overlain in part by lava and to the east by gravell beds at the base of the mesa sandstone. The principal material is soft, yellowish sandstone, in part containing some clay. The top beds have been eroded and the base is

not revealed, although probably not far below the bottom of the valleys.

In the Arroyo San Gregorio, about 6 miles southwest of La Purisima, the yellow beds are exposed lying on Monterey beds and overlain by mesa sandstone. They thin rapidly to the north where the surface of the Monterey beds rises rapidly. The yellow beds outcrop prominently again near Purisima Vieja, 10 miles northwest of La Purisima, exhibiting all the strata down to the richly fossiliferous limestone bed which occurs at the base of the formation near La Purisima. The following section is exposed in this vicinity. Probably Monterey beds are not far below the bottom of the arroyo at this place.

The stratigraphic relations of the yellow beds are extensively exposed in the Arroyo de la Purisima from near its mouth to a point

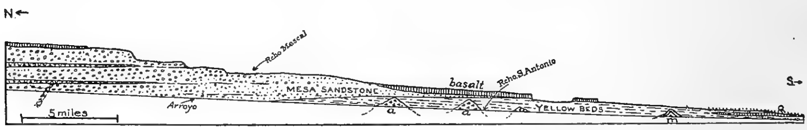


FIG. 15.—Section in the Arroyo San Raimondi northwest of La Purisima, Baja California. *a*, agglomerate; *m*, Monterey beds; *q*, Quaternary.

about 6 miles above La Purisima where the base of the overlying mesa sandstone crosses the canyon. The principal features are shown in the cross-section 14, Figure 2, and the columnar sections in Figure 17. For some distance near the 1,360-foot boring the yellow beds may either thin out or give place horizontally to a massive bed of conglomerate which lies on the Monterey beds; the precise relations are obscured by talus which breaks the continuity of the outcrops. The view, Figure 18, shows high cliffs 10 miles below La Purisima where the yellow beds include a massive member of impure limestone filled with fossils, apparently the same bed as the first one rising above tidewater several miles southwest (see Fig. 17, sec. 1). This bed appears either to thin out or to give place to sandstone and agglomerate farther north near the drill hole. There are extensive exposures of yellow beds about La Purisima (see Fig. 19) with a basal member of limestone filled with

fossils. This member is conspicuous in the bed of the arroyo a short distance below the village where it is crumpled in a series of small but closely appressed flexures. It also outcrops almost continously for several miles below this place, mostly in the bed of the arroyo. At two localities about 4 miles below La Purisima, small arches reveal the top of the underlying Monterey beds with contact clearly exposed. The yellow beds and this basal limestone disappear a short distance farther downstream or near the 1,360-foot boring, as shown in Figure 17. The precise stratigraphic conditions at this place could not be ascertained, but the basal limestone, at least, appears to abut against the old slope of a mound of Monterey beds, probably a local shore line.

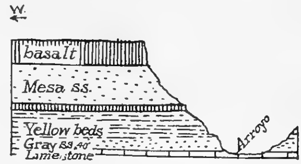


FIG. 16.—Section on the west side of the Arroyo San Gregorio opposite Purisima Vieja, ten miles northwest of La Purisima, Baja California, looking north.

The igneous rocks included in the yellow beds are the products of contemporaneous volcanic action, and it is not unlikely that to the east they grade into the lower part of the great succession of

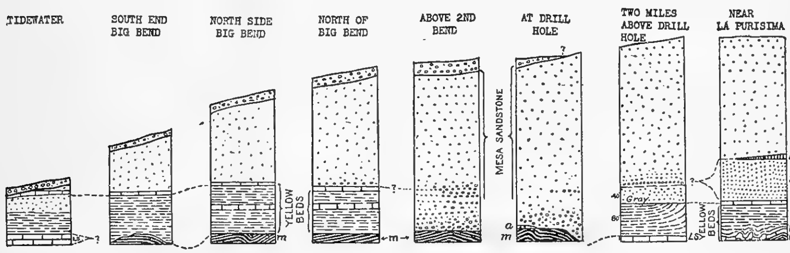


FIG. 17.—Columnar sections along the Arroyo de la Purisima, showing the stratigraphical relations of the yellow beds. *m*, Monterey beds; *a*, agglomerate; *ls*, fossiliferous limestone at base of yellow beds; *s*, massive sandstone at La Purisima; *b*, sheet of basalt in mesa sandstone.

agglomerates, etc., which constitute the base of the high sierra. In the valley of the Arroyo San Raimondi (or San Miguel), 40 miles northwest of La Purisima, there are many exposures of large bodies of agglomerate included in or displacing the upper members of the yellow beds (see Fig. 20). One notable outcrop is near the Rancho San Antonio, where the rocks have the relations shown in

Figure 15. Another similar mass is exposed in the Arroyo Valle, 8 miles west of Rancho San Antonio, and on the Arroyo Vaca, 50 miles west by north of La Purisima, there is an exposure high in the canyon walls showing the relations indicated in Figure 21. This section shows that the agglomerate was erupted during the time of deposition of the yellow beds and prior to the lava flow which now caps the mesas. A small amount of agglomerate lies on the Monterey beds in the Arroyo San Gregorio, 11 miles W. 10° S. of La Purisima, and also near the 1,360-foot bore hole on the Arroyo de la Purisima.

FIG. 18.—Yellow beds and overlying Mesa sandstone on the Arroyo de la Purisima, ten miles below La Purisima, Baja California. Massive fossiliferous bed near center.

Thin sheets of basalt are included in the yellow beds in the lower part of the canyons of the Arroyo San Raimondi and the Arroyo Caije. The relations of the old lava flows are well exposed just south of Caije and along the west side of San Juanico Bay, 7 miles due west of La Purisima. The exposure near Caije is due to a low dome, and here the lava is overlain by a 30-foot, soft, gray sandstone, a local member of the yellow beds. A thin sheet of basalt caps the yellow beds at Purisima Vieja, as shown in Figure 16, and halfway between that place and Poza Honda an igneous mass occupying the bottom of the valley was probably a vent from which this sheet was erupted.

Fossils.—The yellow beds contain fossils at various places which indicate later Miocene age, Carrizo Creek horizon. Possibly the sediments represent a still longer epoch. Fossils collected from the prominent limestone bluff at the head of tidewater at the mouth of the Arroyo de la Purisima were determined as follows by Dr. Julia A. Gardner, of the United States Geological Survey: *Scutella*

FIG. 19.—La Purisima from the south. Pilon in center is Mesa sandstone capped by basalt; slopes of yellow beds below. Lava-capped mesas in the distance.

FIG. 20.—Conglomerates, agglomerates, tuffs, and igneous flows near San Miguel, fifty miles northeast of La Purisima, Baja California.

gabbi Remond: *Terebra* sp., cf. *T. simplex* Carpenter; *Conus* sp., cf. *C. vittatus* Hwass; *Oliva* sp. aff. *O. angulata* Lamarck; *Oliva* n.sp. aff. *venulate* Lamarck; *Phos.* sp.; *Turritella* sp., cf. *T. cooperi* Carpenter; *Turritella* n.sp., cf. *T. ocoyana* Conrad; *Turritella* n.sp. aff. *T. goniostoma* Valenciennes; *Turritella* sp., cf. *T. ineziana* Conrad; *Cancellaria* n.sp., cf. *C. vetusta* Gabb; *Macron merriami* Gabb ?; *Natica pablonesis* Clark ?; *Chlorostoma* (*Omphalius*) sp.

aff. *C. dalli* Arnold; *Glycimeris* n.sp. (2); *Arca* n.sp., cf. *A. samuloensis* Osmont; *Arca* sp. ind.; *Ostrea veatchii* Gabb; *Pecten* (Pecten) *carrizoensis* Arnold; *Pecten* (*Lyropecten*) *crassicardo* Conrad; *Pecten* n.sp. A.; *Cardium* sp.; *Phacoides* sp.; *Mytilus trampaensis* Clark?; *Chione* sp. *Chione* sp., cf. *diabolensis* Clark; *Semele* n.sp. and *Balanus concavus* Brönn. The relations of this limestone are shown in the sections of Figures 1 and 2. Apparently it is at the base of the formation, and the same bed appears again along the arroyo a mile above the 1,360-foot bore hole and outcrops almost continuously to beyond La Purisima. In the vicinity of the latter place Dr. Kew collected the following fossils, determined by Dr. Julia A. Gardner, of the United States Geological Survey: *Strombus* n.sp.; *Turritella* n.sp.; *T.* n.sp., cf. *T. ocoyana* Conrad; *T.* sp., cf. *T. margaritana* Normand; *Calyptraea costellata* Conrad?; *Arca* sp. *Ostrea* sp. ind. *Pecten* n.sp.; *Venericardia* sp., cf. *V. californica* Dall; *Chione* sp. *Chione* n.sp.: *Balanus concavus* Brönn—a Carrizo Creek fauna.

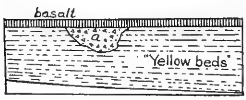


FIG. 21.—Section in east wall of canyon of the Arroyo de la Vaca, showing relations of agglomerate in the yellow beds.

The higher limy member occurring in a thick ledge near the middle of the yellow beds along the walls of the Big Bend and for some distance above it along the Arroyo de la Purisima, as shown in Figures 17 and 18, contain many fossils. A large collection was made from this horizon, but it has been misplaced in the National Museum in Washington. The basal limestone of the yellow beds outcropping in the bed of the Arroyo San Gregorio near Purisima Vieja yielded the following forms: *Purpura*, cf. *P. vaquerosensis* Arnold var.; *Turritella*, cf. *T. ocoyana* Conrad; *T.*, cf. *T. ineziana* Conrad; *T.*, cf. *T. margaritana* Normand: *Arca* sp. ind. *Pecten* sp., cf. *P. estrellanus* Conrad; *Pecten crassicardo* Conrad; *Pecten* sp., cf. *Pecten* n.sp., *Cytherea* n.sp. ind.—a Carrizo Creek fauna.

Only a few fossils were found in the extensive exposures of yellow beds north of the Arroyo Valle, 45 miles northwest of La Purisima. In a lower sandy member are abundant *Ostrea bourgeosii* Gabb and *Turritella*, cf. *T. jewetti*. At another place were collected, from the middle beds, *Chione*, cf. *C. elsmerensis* English and many specimens of *Tellina*, cf. *T. ocoides* Gabb. These fossils

were determined by Dr. Julia A. Gardner, who regards them as probably Upper Miocene.

Fossils collected near the top of Cerro Angel, 18 miles west of San Ignacio, were identified as follows: *Oliva* sp.; *Turritella*, cf. *hoffmani* Gabb; *Calyptraea*, cf. *diegoana* Conrad; *Glycymeris* sp.; *Arca* sp.; *Pecten* n.sp.; *Pecten* sp.; *Dosiniaarnoldi* Clark; *Chione elsmerensis* English?; *Chione* sp.; and *Macoma vaulecki* Arnold. They are regarded as San Pablo or late Miocene.

These yellow beds outcrop extensively under the lava mesas about San Ignacio (see Fig. 14) and in the arroyo below that village. Fossil oysters are abundant in some of the layers.

Yellow beds similar to those of the La Purisima region overlie Monterey beds in the slopes east of Rancho Tepetate. The steeply uplifted strata just northwest of Rancho Platana include a richly fossiliferous limy ledge which yielded fossils of the same fauna as those occurring in the basal limestone member of the yellow beds in the Arroyo de la Purisima. The following were identified by Dr. Julia A. Gardner: *Strombos* n.sp. (same as one at La Purisima); *Turritella* n.sp. 1 and 2; *Codakia* sp.; *Cardium* sp. ind.; *chione* sp., cf. *C. fernandoensis* English; *C.* sp. ind. and *Psephidea?* sp.—a Carrizo Creek fauna. It is reported that later Miocene beds outcrop on the gulf shore near Cayote Point and Agua Verde Bay, doubtless coming up from under the agglomerate series, as shown in section 4.

From a bed in the base of the formation or not far below it, 8 miles northwest of El Pilon, were collected the following fossils: *Turritella* sp., cf. *T. margaritana* Normand; *Turritella* N. sp.; *Yoldia* sp.; *Ostrea* sp. ind.; *Mytilus?* sp.; *Modiolus?* sp.; *Pecten?* sp., *Chione* sp., cf. *C. fernandoensis* English and *Mactra?* sp.; determined by Dr. Julia A. Gardner who regards them as late Miocene or early Pliocene.

MESA SANDSTONE AND THE GREAT LATER TERTIARY VOLCANIC SERIES

Gabb recognized the fact that the widespread mesas of southern Baja California consist of a thick mass of west-dipping sediments and conglomerates. He named it the "mesa sandstone," and while he erroneously extended the application of the name to other

formations constituting mesas in the central and northern parts of the peninsula, the name may be useful in the southern part of the region until a more definite classification is practicable. I find that the formation presents two phases: a massive, gray sandstone, several hundred feet thick in the western portion of the mesa region, rapidly merging into conglomerates with thick bodies of agglomerate and tuff to the east. This relation is shown in sections 11–20, Figures 3 and 4, and is a most striking feature. The coarse sediments are in hard, massive beds, 4,000 feet thick in places, constituting the high sierra extending continuously southward

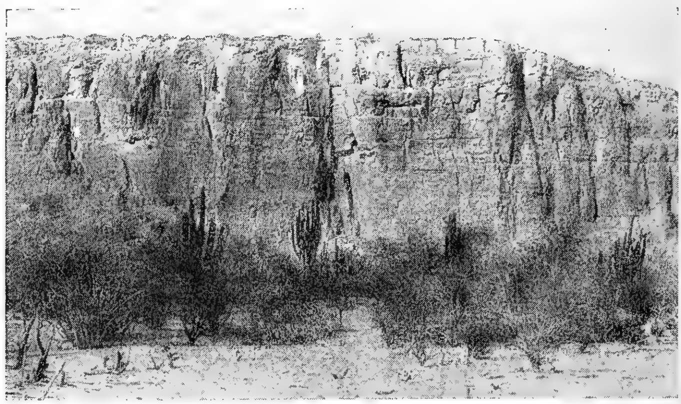


FIG. 22.—Mesa sandstone on the Arroyo San Andres, twenty-one miles south of Comondu, Baja California. Shows massive bedding.

from near latitude 28° to beyond La Paz. Sheets of contemporaneous igneous rocks are included and the succession is penetrated by many intrusions, notably in the Mulegé and Santa Rosalia regions. To the east it lies on schist, granite, etc., and to the west on sandstone of earlier Tertiary age. South of the latitude of La Paz the thickness diminishes, volcanic rocks are not present, and at Todos Santos the underlying rocks reach the shore of the Pacific Ocean. I did not observe its relations north of latitude 28°.

In the area of the great, lava-covered mesas extending along the Pacific slope past San Ignacio, La Purisima, and Comondu, the yellow beds are overlain by massive, gray, mesa sandstone. This sandstone is conspicuous in the walls of many canyons which

cut the mesa zone, notably in the arroyos San Raimondi (or Miguel), San Gregorio, De la Purisima, Comondu, San Benancio, San Andres, and Santo Domingo. The greatest thickness observed in this area is 300 feet; to the west the beds are thin out to nothing under the lava caps of the mesas, and to the east they merge into the great formation of detrital and volcanic rocks constituting the central sierra. This transition is well exposed in the Arroyo San Raimondi near the little pueblo of San Miguel, in the canyon of the Arroyo San Gregorio above Purisima Vieja, in the Arroyo de la Purisima above Huerta Vieja (a few miles above La Purisima), and it is also evident in the canyons from the Arroyo Comondu southward. The change to the east consists in the sediments becoming coarser and thicker, and probably overlying beds also are present under the great sheets of the basalt which cover the high mesas. Close scrutiny would probably reveal the presence of several formations in the sandstone series. A feature of this kind is clearly apparent about La Purisima where a harder, massive sandstone appears as a huge wedge in the other sediments (see Fig. 17). It thins out to the southwest, a mile or two below La Purisima.

The eastern extension of the mesa sandstone was not studied in detail, but was observed in various trips across the high sierra. The predominating rock is conglomerate in thick beds and consisting largely of bowlders of igneous rocks together with some quartzite. Sheets of lava, layers of tuff, and thick bodies of agglomerate are included. Some of the lavas are basalt, others are light-colored, more acidic rocks. To the east many intrusive masses are present. These are conspicuous west of Santa Rosalia, near Mulegé, east of Conception Bay, and southward past Loreto.

A fine exposure of mesa sandstone on the west bank of the Arroyo San Gregorio, 10 miles slightly north by west of La Purisima, shows the following strata:

	Feet
Lava sheet at top of mesa	00+
Sandstone, gray, massive, compact	60
Sandstone, white, soft	120
Bowlders and sand	40
Agglomerate, reddish in places	4
Yellow beds in bottom of the arroyo	30

Near Purisima Vieja, on the Arroyo San Gregorio, 12 miles northwest of La Purisima, a thin lava bed lies at the base of the formation (or in the top of the yellow beds, see Fig. 16), and a short distance above that place is a vent from which this lava flow probably came. Near San José, 7 miles farther up the same canyon, conglomerates predominate in the high canyon walls and two sheets of basalt are included. The later are well exposed in Arroyo Caije near Rancho Nombre de Marie, 25 miles south-south west of Mulegé. In the extensive exposures of the agglomerates and conglomerates near San Miguel on the Arroyo Raimondi (or Miguel), as shown in Figures 15 and 20, a thin sheet of trachyte is included in the succession. Near Rancho los Angeles in the same canyon the cliffs 500 feet high consist of coarse conglomerate, some beds being harder than others and consisting largely of bowlders of igneous rocks in gray sand. Below Rancho Mescal, a few miles above Rancho las Tules, underlying gray sandstone appears lying in yellow beds and in places on agglomerate which displaces the upper members of the yellow beds. These features are also well exposed halfway between Rancho las Tules and Rancho San Antonio. Not far below the latter place the basalt beds of the mesa sandstone thin out and the lava of the mesas lies directly on top of the yellow beds. The relations in this arroyo are shown in Figure 15.

Mesa sandstone appears at San Vicente, a ranch 5 miles southeast of La Purisima, where a fine spring issues from it, and is well exposed along the arroyos San Antonio and Pabellon 10–20 miles south of La Purisima. In the latter there are high bluffs of it at Rancho Pabellon and extending to beyond Coyote hole a watering-place as head of tidewater, 20 miles south-southwest of La Purisima. Here the rock is soft sandstone, in part pale greenish and yellowish. It is capped by caliche and conglomerate of Quaternary age which covers the adjoining mesas. The yellow tint may be due to a mixture of detritus from yellow beds which do not outcrop in this vicinity.

In the deep canyon of the Arroyo Comondu, near the village of Comondu, there are high walls of mesa sandstone capped by basalt of the lava sheets which cover the adjoining mesas, and there is also a cap of basalt at lower level, apparently a remnant

of a sheet which flowed down the canyon when it had only about half its present depth. It is possible, however, that this lava sheet is included in the mesa sandstone, for such a relation is shown on the south side of the canyon 12 miles below Comondu, just above the mouth of the Arroyo Belamote. In that vicinity also there are some conglomerate members showing strong unconformity at base, and a few conglomeratic layers are included in the gray sandstone. It is probable that the mesa-sandstone outcrop extends down the Arroyo Comondu to Rancho San Andreta, near the mouth of the arroyo, where the following section is exposed:

Conglomerate	30 feet, many fossils
Gray sand, pale greenish tint	20 feet
Limy ledge, yellowish	5 feet exposed

Fossils collected from the basal bed in this exposure were determined by Dr. Gardner as *Turritella, Cytherea* and *Balanus* ?, and a *Chione* resembling *C. elsmerensis* English, probably Pliocene in age. This is the only paleontologic evidence that I obtained as to the age of the formation and it is not conclusive.

Extensive exposures of mesa sandstone are presented in the Arroyo San Benancio and the Arroyo San Andres south of Comondu, one of which is shown in Figure 22. The sandstone, conglomerate, agglomerate, tuff, etc., constitute cliffs and high canyon walls along the west side of the Gulf of California from Agua Verde Bay to Canyon de los Reyes. A section in the latter shows the following features:

GENERALIZED SECTION OF STRATA IN CANYON DE LOS REYES

	Feet
Soft, gray sandstones, some beds conglomeratic (top of ridge)	220
Rhyolite flow (bench-maker)	40
Soft, gray sandstone, several beds of conglomerate to the east	260
Conglomerate with bowlders of igneous and other crystalline rocks	40
Rhyolite flow (making wide bench)	30–70
Gray conglomeratic sandstone with agglomerate layers	200
Agglomerate, tuff, volcanic ash, igneous sheets. Bedding massive extends to tide-level	650
Total thickness	1,650

The thin sheets of rhyolite included in this section thin out to the south, but thicken considerably to the north, and the mass of agglomerate also thickens greatly in that direction. This series appears again in ridges passing just east of La Paz and on the islands of Espiritu Santo, Partida, and San Josef, where it consists of thick bodies of agglomerate, tuff, and eruptive rocks. The thinning and fining of the formation to the west is plainly visible in many canyons on the west slope of the sierra west and northwest of La Paz as far north as the Arroyo Santa Cruz near latitude 25° 20′. In this part of the peninsula the soft, gray, massive, typical "mesa sandstone" appears extensively in the highlands of the sierra as a component of the complex; but to the west, as the beds thin and fine, it becomes the dominant feature. It is exposed in the arroyos Conejo, Datilar, Guadalupe, 35–60 miles west of La Paz, and in the cliffs just below San Hilario; but at all of these places layers of pebbles are included and scattered bowlders occur. About El Pilon, 10 miles north of Hilario, it appears in many cliffs, some of which show limy layers. Farther north the conglomerate admixture increases notably in the vicinity of Rancho Jesus Maria, 35 miles north-northwest; but at the north side of Cerro Nombre del Dios, 6 miles southwest of that ranch, the gray sandstone contains only a few thin beds of conglomerate and widely separated bowlders. The westernmost exposure observed in this portion of the peninsula was on the Arroyo San Luis, 17 miles northwest of San Luis, where the sandstone is fine-grained and massive.

PLIOCENE TO POST-PLIOCENE

Along the wide belt of lower lands adjoining the Pacific south of latitude 26° 30′, there is a cover of sand and gravel with limestone members which doubtless include not only the Quaternary, but probably also a formation of Pliocene age. These deposits vary in thickness from a few feet to 200 feet or more, and they extend far up the slopes, and on the higher mesas they are probably represented by a thick cap of bowlders. The latter are especially well exhibited on the high mesas north and northeast of San Hilario and on the long sloping mesas crossed by the main trail north from San Luis. The wide, low plains of the Magdalena region are

covered by a deposit of sand of Quaternary age which to the south extends to the Arroyo Salado and beyond, and some distance up the slopes to the east. At Aqua Verde near Rancho Salado (latitude 24° 31′, longitude 111° 31′), underlying yellow sands with hard, limy ledges appear in a cliff 40 feet high. The section at this place is as follows:

	Feet
Gray conglomerate sandstone, which floors the adjoining low sloping mesa	15
Yellow, fine-grained sandstone somewhat harder beds above, and at base a wedge-shaped, hard, limy ledge with fossils	25

The fossils collected here were determined by Dr. Julia A. Gardner as follows: *Turritella* sp.; *Arca* aff. *A. microdonta* Conrad; *Mytilus* sp.; *Modiolus* sp.; *Periploma* (*Halistrepta*) *sulcata*, Dall?; *Tivela* n.sp.; *Chione latilamenosa*, A and M? *C.* sp.; *Metis* aff. *M. alta* Conrad and *Balanus* sp. "Age post-Miocene."

These yellowish beds extend for 3 or 4 miles up the Arroyo Salado and lie on the Eocene (?) sandstone. Probably their yellow color is due to material from the Monterey formation or yellow beds underlying them. They extend up the adjoining plateaus to an altitude of 500 feet or higher. Limestones of this formation are conspicuous in the Cerrito Flor de Melba, 60 miles west of La Paz, and on the walls of various arroyos from Datilar to Cuaño, as well as in cliffs along the ocean (60–40 miles west of La Paz). These limestones occur at several horizons, and they are highly fossiliferous at most places, but the fossils of the upper beds at least appear to be post-Pliocene. The dip is to the southwest at a low angle, and the beds extend far up the west slope of the sierra where they overlie the mesa sandstone.

THE GREAT UPLIFT

It has been found that much of the peninsula of Baja California has been uplifted out of the sea in very recent geologic times. Deposits of modern sea shells occur at many places in regions up to altitudes of 1,000 feet and they are reported as high as 3,300 feet. Old belts of sand dunes and shore lines are conspicuous far above

sea-level in several areas. Wittich[1] has presented many interesting facts relating to this subject with illustrations of shell deposits and shore features up to altitudes of 1,052 meters in a divide near Mission San Borja; this was as high as he ascended in his explorations.

The emergence has occupied a somewhat long time and may still be in progress. The shore erosion has been effected at various stages, although some of it may have been developed during the preceding submergence. The latter was apparently relatively transient, and most of the present configuration of the land was developed prior to this event.

[1] Ernesto Wittich, "La Emersion Moderna de la Costa Occidental de la Baja California," *Soc. Cien. Antonio Alzate (Mexico) Memoirs*, Vol. XXXV, pp. 121–44.

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